HYDROLOGICAL REGIME CHANGES IN A CANADIAN PRAIRIE WETLAND BASIN

A Thesis Submitted to the

College of Graduate Studies and Research

In Partial Fulfillment of the Requirements

for the Degree of Master of Science

in the Department of Geography and Planning

(Centre for Hydrology)

University of Saskatchewan, Saskatoon

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Preface

The research presented in this thesis is partially published in the following article:

Dumanski S, JW Pomeroy, CW Westbrook (2015) Hydrological regime changes in a Canadian Prairie wetland basin, Hydrological Processes Special Issue, DOI:10.1002/hyp.10567.

I led the publication with the mentoring support of my co-advisors, who co-authored the paper. The paper is not presented verbatim here; rather, I have provided considerable more detail and data analysis. In addition, some components of this thesis are reported in a non-peer reviewed report targeted at one of the funding agencies:

Pomeroy, J.W., K. Shook, X. Fang, S. Dumanski, C. Westbrook, and T. Brown (2014) Improving and testing the Prairie Hydrological Model at Smith Creek Research Basin, Centre for Hydrology Report No. 14, May 2014.

Abstract

The hydrology of the Canadian Prairies has been well described in the scientific literature. 20th C observations show that snowmelt over frozen soils accounted for over 80% of the annual runoff, and streamflow hydrographs peaked in April and ceased in May due to a lack of runoff or groundwater contributions. Since then, the region has undergone rapid changes in land use and climate, both which affect streamflow generating processes. This study evaluates the detailed hydrological impact of regional changes to climate on an instrumented research catchment, the Smith Creek Research Basin (SCRB); an unregulated, wetland and agriculture dominated prairie catchment in south-eastern Saskatchewan. Wetlands have been drained for decades, reducing wetland extent by 58% and maximum storage volume by 79%, and increasing drainage channels lengths by 780%. Long term meteorological records show that there have been gradual changes to the climate: though there are no trends in annual precipitation amount, increasing temperatures since 1942 have brought on a gradual increase in the rainfall fraction of precipitation and an earlier snowmelt by two weeks. In the summer months, the number of multiple day rainfall events has increased by 5 events per year, which may make rainfall-runoff generation mechanisms more efficient. Streamflow records show that annual streamflow volume and runoff ratios have increased 14-fold and 12-fold, respectively since 1975, with major shifts in 1994 and 2010. Streamflow contributions from rainfall-runoff and mixed-runoff regimes increased substantially. Snowmelt runoff declined from 86% of annual discharge volume in the 1970's to 47% recently while rainfall runoff increased from 7% to 34%. Annual peak discharge tripled over the period from 1975 to 2014, with a major shift in 1994, while the duration of flow doubled in length to 147 days after a changepoint in 1990. Recent flooding in the SCRB has produced abnormally large streamflow volumes, and flooding in June 2012 and 2014 was caused solely by rainfall, something never before recorded at the basin. Although the observed changes in climate and wetland drainage are substantial, it is unlikely that a single change can explain the dramatic shifts in the surface hydrology of the SCRB. Further investigation using process hydrology simulations is needed to help explain the observed regime changes.

Acknowledgements

This project would not have been possible without the help of many individuals. Firstly, I would like to thank my supervisors Dr. John Pomeroy and Dr. Cherie Westbrook for their expertise and guidance. The encouragement and invaluable feedback received were fundamental to the success of this project. I would also like thank my thesis committee, Dr. Chris Spence, and external examiner, Dr. Saman Razavi, for their time and helpful suggestions.

Secondly, I would like to thank those who helped with field work and technical support. A great amount of appreciation goes out to Dr. Kevin Shook for the countless hours spent sharing knowledge and providing support, and to Lyle Boychuk for sharing information and lending a helpful hand. As well, Logan Fang, Nicole Seitz, Joni Onlcin, and Bryan Page were invaluable for all of their help with data and last minute trips out to the field.

To my fellow graduate students: thank you for the countless hours of help and guidance throughout my thesis. I was extremely fortunate to have been part of such a supportive group of peers, including Chris, Jen, Emily, Phillip, Evan and many others.

This project would not have been possible without the funding support from the University of Saskatchewan, Ducks Unlimited Canada, and the Canada Research Chairs program. Also, thank you to the contributors to R (software environment for statistical computing and graphics) which was used to conduct all of the analysis in this thesis.

Lastly, I want to thank my friends, family, and spouse, Jeremy, for the many years of love and support. I am forever grateful for the unwavering encouragement throughout all of the highs and lows. I could not have finished this work without it.

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List of Abbreviations

AHCCD	Adjusted and Homogenized Canadian Climate Database
CWCS	Canadian Wetland Classification System
DEM	Digital Elevation Model
ENSO	El Niño Southern Oscillation
ET	Evapotranspiration
PDO	Pacific Decadal Oscillation
PPR	Prairie Pothole Region
SCRB	Smith Creek Research Basin
SOI	Southern Oscillation Index
SWE	Snow Water Equivalent
w.e.	Water Equivalent
WSC	Water Survey of Canada

Chapter 1: General Introduction

1.1 Introduction

The hydrology of the Canadian Prairies has been described as a semi-arid, cold regions system where snowmelt runoff over frozen soils dominates streamflow generation, substantial runoff is stored in wetlands resulting in variable basin contributing areas and summer rainfall primarily supplies crop evapotranspiration (Gray, 1970; Gray et al., 1986; Granger and Gray, 1990; Granger and Gray, 1989; Pomeroy et al., 2007; Fang et al., 2010). But re-assessment is necessary as the region has undergone changes in climate (Akinremi et al., 1999; Millet et al., 2009; Bonsal et al., 2012; Shook and Pomeroy, 2012) and land use (e.g. Rashford et al., 2011) which impact streamflow generating processes (e.g. Burn et al., 2010).

There has been considerable research describing changing climate or evolution of land use in the region, with few studies examining the cumulative impact that these changes have had on streamflow. These few studies have been conducted on large river basins and have resulted in mixed conclusions. For example, Miller and Nudds (1996) examined 12 unregulated rivers from Canada and U.S. and found that landscape alteration, not changes in precipitation, caused an increase in runoff. In contrast Ehsanzadeh et al. (2012, 2014) could not find a detectable impact of climate change, wetland drainage or farming practices on streamflow frequency distributions in the Canadian Prairies from a range of rivers that included regulated systems. There is thus a need for a detailed analysis of the changes in climate, wetland drainage, and runoff processes to better understand the dimensions of the hydrological change in the region.

1.2 Thesis Goal and Objectives

The goal of this thesis is to improve the understanding of the historic variability of the hydrometeorology, runoff processes and land use in the Canadian Prairie Pothole Region. Owing to the lack of detailed regional data, a case study was conducted in a well-documented, wetland dominated headwater basin in the Canadian Prairies, the Smith Creek Research Basin. The objectives of this study are to:

1. Describe the historical variability of select hydroclimatic variables
2. Evaluate the hydrological change in the context of climate and land use change

Chapter 2 : Literature Review

This section reviews the current state of the knowledge of the climate, hydrology and land use of the Prairie Pothole Region (PPR) which spans three Canadian provinces and four American states. Included is an analysis of the literature that has examined observed and modelled changes in each of the topics. This section begins with a description of the climate and changes in climate observed over the last century, followed by a depiction of the unique hydrological setting of the PPR. Changes in land use are discussed as well as the influence that such changes in climate and land use have had on the hydrology of the region.

2.1 Late 20th Century Climate Setting of PPR

PPR is located in mid-central North America and extends over 750 000 km^2 (Millet et al., 2009). The Canadian portion of the PPR is a cold region and is characterized by long, cold winters with continuous snowcover and frozen soils throughout most of the region (Pomeroy et al., 2009). Average annual precipitation is typically less than 500 mm/year (Burn et al., 2008), with snowfall accounting for about one third of that (Gray & Landine, 1988). In the spring, snowmelt typically occurs over frozen soils, resulting in high amounts of runoff due to limited or restricted infiltration (Gray et al., 1995). Although snowfall accounts for approximately one-third of the annual precipitation, over 80% of the annual surface runoff is sourced from snowmelt over frozen soils (Gray and Landine, 1988).

Rainfall events throughout the spring and early summer are typically from frontal systems, whereas summer precipitation is usually from convective storms supplying intense rainfalls over small areas (Gray, 1970). Rainfall predominantly falls as small events (0.5 – 5 mm), yet these small events account for less than 25% of annual rainfall (Akinremi et al., 1999). A majority of the rainfall occurs from mid-June to early July (Bonsal et al. 1999), most of which is consumed by high rates of evapotranspiration (Armstrong et al., 2008), quickly depleting soil moisture levels. Therefore, rainfall events tend to produce zero to minimal runoff during the summer months (Granger and Gray, 1989).

Using a modelling approach, the estimate of total evaporation during the growing season on the Canadian Prairies ranged from 150-200 mm during drought conditions to over 400 mm when moisture conditions were above normal with daily estimates ranging from 2.8 mm/day to over

5.0 mm/day (Armstrong et al., 2015). The annual potential evapotranspiration (ET) rates are much higher and range from 600 – 900 mm/year on average (Burn et al., 2008).

Owing to the mid-continental location, long distances from large water bodies, high potential ET and high precipitation variability, the agricultural regions of the Canadian Prairies are prone to droughts (Bonsal et al., 2013). Droughts are frequent, with almost every decade having at least one drought year since records started in the 17th century (Nkemdirim and Weber, 1999). Periods of droughts can last from a single year up to many years. The drought of 1961 was the worst single year drought on record (Bonsal et al., 1999), while the drought of 1999-2004 was the most severe multi-year drought (Bonsal and Wheaton, 2005). Precipitation deficits and reduced soil moisture during droughts have important hydrological impacts on both streamflow and the replenishing of water bodies (Nkemdirim and Weber, 1999; Fang and Pomeroy, 2007).

2.2 Changes in Climate

Owing to the semi-arid climate where potential ET exceeds precipitation, the hydrology of the PPR is highly sensitive to climate change, as well as land use changes (Conly and van der Kamp, 2001; Fang and Pomeroy, 2007). Climate plays an important role in the amount of water that enters or leaves a wetland (Millet et al., 2009). Many studies have looked at how climate has changed throughout Canada (Zhang et al., 2000; Vincent et al., 2007; Mekis and Vincent, 2011). Some of these studies have focused attention on the Canadian Prairies (e.g. Akinremi et al., 1999; Millet et al., 2009; Shook and Pomeroy, 2012; Bonsal et al., 2013).

Across Canada, mean annual temperatures have increased between 0.5 °C to 1.5 °C, with the average increase of 1 °C during the second half of the 20th century (Zhang et al., 2000; Vincent et al., 2007). Comparable changes in temperature have been noted in the Canadian Prairies. Millet et al. (2009) found that maximum temperatures have generally declined while minimum temperatures have increased. Significant increases in minimum temperatures were observed in winter and summer, with greater increases in winter. Maximum temperatures were found to be increasing in winter, yet cooled in the summer. In contrast, Zhang et al. (2000) showed that from 1900 to 1999, both minimum and maximum temperatures have increased on the Prairies, with minimum temperatures increasing more than maximum. Annual mean daily minimum temperatures increased 2.5°C, with spring minimum temperatures warming the most (no value given). During the same time period, annual mean daily maximum temperatures increased 1.5°C on the Prairies, with warming being greatest in spring (2°C). Similar to Zhang et al., (2000), Bonsal et al. (2001) also concluded that the greatest increases in minimum and maximum temperatures occurred in winter and early spring from 1950 to 1998, and only minimum temperatures were found to have increased in summer. Most researchers agree that enhanced continental interior drying will be brought on by increases in temperatures (e.g. Bonsal et al., 2013), and may exacerbate droughts during the 21st century.

Precipitation throughout Canada, including the prairie region, was generally found to be increasing (Akinremi et al., 1999; Zhang et al., 2000; Fernandes et al., 2007; Vincent, 2007; Millet et al., 2009; Mekis and Vincent, 2011). Across the PPR, precipitation increased by about 9% during the 20th century (Millet et al., 2009). Similar results were presented in Akinremi et al. (1999) who found precipitation was increasing by 0.62 mm/year on the Canadian Prairies between 1920 and 1995. A general consensus between most studies found that increases in precipitation were mainly due to increases in rainfall (Akinremi et al., 1999; Vincent, 2007; Mekis and Vincent, 2011) at a rate of 0.6 mm/year from 1920-1995 (Akinremi et al., 1999) while snowfall has significantly decreased (Akinremi et al., 1999; Mekis and Vincent, 2011). Particularly, precipitation in the spring and fall was found to be falling more as rainfall than snowfall over the 20th century across the Canadian Prairies (Shook and Pomeroy, 2012).

At many sites across the Canadian Prairies, the number of multi-day rainstorms has increased significantly, while the number of single day rainstorms has decreased from 1951-2000 (Shook and Pomeroy, 2012). This is important for runoff generation as multi day rain events tend to be frontal in nature and have the ability to produce runoff at larger scales (Hayashi et al., 1998; Shook and Pomeroy et al., 2012). Additionally, the intensity of the single day rainfall events has declined (Akinremi et al., 1999; Shook and Pomeroy, 2012).

2.3 Topography and Surface Hydrology

Due to the millions of internally drained depressions on the landscape and the semiarid climate, the hydrology of the PPR is distinctive. During the last deglaciation, tens to hundreds of meters of clay-rich glacial till was deposited (Lennox, 1988). As the Wisconsin glacier receded, large amounts of small, stagnant ice blocks separated and disintegrated forming the “knob and kettle”, or hummocky, topography (Gravenor and Kupsch, 1959). Such topography is described in Gravenor and Kupsch (1959: pg. 50) as “a nondescript jumble of knolls and mounds of glacial debris separated by irregular depressions”. The hummocky landscape formed enables surface runoff to internally drain into the depressions, forming wetlands or sloughs. It is estimated that there are 1.5 million wetlands in the agricultural portion of Saskatchewan, with 80% of them having a surface area of < 1 ha area (Huel, 2000).

The water balance of wetlands is controlled by interactions between precipitation, evapotranspiration, overland flow, and subsurface flow (Winter, 1989; Winter and Rosenberry, 1995; van der Kamp and Hayashi, 2009). The seasonal wet/dry periods cause fluctuations in wetland water levels (Woo and Rowsell, 1993) due to the high potential ET that exceed precipitation (Burn et al., 2008). The main source of water to wetlands on the prairies is snow (Gray and Landine, 1988; Covich et al., 1997; Fang and Pomeroy, 2008) and is critical to the existence of wetlands (LaBaugh et al., 1998). Throughout the winter months, snow is redistributed by wind from open or exposed sites to sheltered areas, including vegetated areas, depressions and stream channels (Fang et al., 2009). Snow redistribution to vegetation and topographic depressions (Pomeroy et al., 2007) results in a heterogeneous snowcover. In the spring during snowmelt, the snowmelt water infiltrates into the underlying soil, evaporates, or runs off (Gray et al., 1986). The PPR is in a cold region, such that the soils freeze during the winter, which limits the amount of infiltration through the formation of ice within the soil pores (Komarov and Makarova, 1973). In the absence of macropores (i.e. cracks), the infiltrability of the frozen soil is regulated by the frozen moisture content of the soil (Gray et al., 1986).

During the summer months, runoff from rainfall events is rare as infiltration rates typically exceed rainfall rates. Soils tend to be unsaturated due to high evapotranspiration rates that quickly deplete soil moisture levels (Shook and Pomeroy, 2012). Rainfall runoff can be produced via saturation overland flow caused by large summer storms (i.e. frontal events) that persist for long periods of time, but these events are rare (Hayashi et al., 1998; Shook and Pomeroy, 2012). Rainfall events tend to be in convective form and have the ability to produce intense rainfall rates. But these convective events vary spatially, cover small areas and do not typically produce enough runoff to cause a change in the discharge (Dyck and Gray, 1979) at a basin scale. A study conducted in the Canadian Prairie provinces found that 79% of 1000 summer rainfall events (>10 mm in 24 hr) from 2000-2004 were solely or partially convective in nature (Raddatz and Hanesiak, 2008).

Water loss from wetlands during the summer is predominately from evapotranspiration (Mills and Zwarich, 1986; Winter, 1999). This includes both direct evaporation from open water and the loss of water through lateral subsurface flow to surrounding plants and leaves the system through evapotranspiration (Mills and Zwarich, 1986; Hayashi et al., 1998). The transpiration by vegetation surrounding the wetland accounts for up to 70% of the water that infiltrates (Parsons et al., 2004). Due to the clay rich glacial sediments throughout the Canadian Prairies (Lennox, 1988, van der Kamp and Hayashi, 1998), deep groundwater aquifers are confined (Nachshon et al., 2014) and recharge has been found to be extremely slow (decadal process; Si and de Jong, 2007) and has negligible effects on the water balance of wetlands (van der Kamp and Hayashi, 2009). Infiltration rates to groundwater range from 1-3 mm/year, or approximately 1% of the annual precipitation (Hayashi et al., 1998, Si and de Jong, 2007).

The minimal interaction between wetlands and groundwater results in the reliance of wetland water supply on precipitation and leaves them vulnerable to changes in climate (Covich et al., 1997). Poiani and Johnson (1993) identified using model simulations that wetland water levels are more sensitive to increases in precipitation (rather than decreases) when temperatures are increased by 2 °C. The study also identified that the influence of precipitation changes on wetland water levels at +4 °C was less than at +2 °C. Through the use of models, Johnson et al. (2005) found that warmer and wetter conditions could counterbalance the effect on the water balance, showing that (in general) a 3 °C increase in temperature could be compensated by a 20% increase in precipitation. Increased frequency and duration of drought conditions could be brought on by increased temperatures and decreased precipitation (Johnson et al., 2005). The effects of increasing temperatures (with a lack of increased precipitation) will affect the smaller, seasonal wetlands (Fennessey, 2014).

The millions of depressions in the PPR are typically geographically isolated and act as closed basins (Hayashi et al., 2003) due to a lack of naturally integrated stream channels (Covich et al., 1997; LaBaugh et al., 1998). These depressional wetlands have a large capacity to store water (Hayashi et al., 2003; Minke et al., 2011). Under normal conditions (1:2 year peak flow), these internally drained areas do not contribute to any stream and are called non-contributing areas (Godwin and Martin, 1975). Storing runoff helps sustain flow in streams and possibly reduce peak flow during smaller floods (Pomeroy et al., 2007). During wetter periods, increased connectivity between depressional wetlands can occur through intermittent surface-water connections (Leibowitz and Vining, 2003) which may result in a temporary increase in contributing area for streamflow.

The size of the contributing area to streamflow is influenced by depressional storage, antecedent conditions and temporal persistence of climatic conditions, resulting in areas that contribute to streamflow one year, but not the next (Stichling and Blackwell, 1957; Ehsanzadeh et al., 2012b; Shaw et al., 2012). In general, the deeper the snowpack and the higher the fall soil moisture content, the larger the area is that contributes to streamflow (Gray et al., 1986). Under snowmelt or wet conditions, the amount of runoff produced can exceed the storage capacity of the depressions, causing water to spill over to a lower-lying wetland through a fill and spill process (van der Kamp & Hayashi, 2009; Brunet and Westbrook, 2012). This process increases the surface connectivity of the basin, and in turn, increasing the streamflow derived from runoff (Fang et al., 2010; Pomeroy et al., 2010). Once the depressions drain below their spill level, the surface outflow ceases. Therefore, the changing connectivity between wetlands results in dynamic contributing areas for runoff (Shaw et al., 2012) and redistributes water to lower lying wetlands. These intermittent streams exist during snowmelt and subsequent flow periods can occur after high rainfall events (Shaw et al., 2012).

The shape and slope of the flow frequency curve is affected by the depressional storage in the watershed, and minor increases in precipitation can result in a disproportionate change in the runoff frequency curve due to an increase in contributing area (Ehsanzadeh et al., 2012b). As water levels within the depressions rise and fall, wetlands expand and contract causing connections to form and break (Ehsanzadeh et al., 2012b). The relationship between contributing area and storage is hysteretic with a sharp decrease in contributing area as depressional storage evaporates or percolates into the subsurface (Shook et al., 2013). Further, Spence (2007) and Shaw (2010) identified a nonlinear relationship between the water storage in the basin and the fraction of a basin that is contributing runoff to the outlet. It is important to note that anthropogenic infrastructure like roadways and culverts have an influence on the contributing areas (Shaw et al., 2012) as they block and restrict flow paths.

2.4 Changes in Land Use

Prior to the European settlement in the early 1900's, the PPR was in a natural state and consisted of grasslands and wetlands (Heagle et al., 2013). Since then, most of the area has been converted to cropland or pastureland (Upper Assiniboine River Basin Study, 2000). The intensification of agriculture is expected to continue in the PPR (Rashford et al., 2011). In association with the increased agricultural area is an increase in the number of dirt and gravel roadways that are organized into a grid pattern. In Saskatchewan, there is 165,000 km of grid roads in which 19,000 km were constructed in the late 1950's in response to the increased rural population and use of automobiles after World War II (Stewart, 2006). These grid roads are elevated approximately 0.5 m to 1.5 m above the ditch (Duke et al., 2003) and act as barriers to overland flow (Pomeroy, 1985; Duke et al., 2003). To artificially allow water to pass through the grid road system (Smith, 1985), culverts have been built into the roadways and range in sizes based on peak discharge and annual precipitation characteristics of the watershed. Some culverts have the ability to regulate the amount of water the passes through a culvert through manually controlled gates situated at the inlet of the culvert that can be cranked up or down (Smith, 1985).

In order to increase agricultural production, efforts have been renewed to drain wetlands (Watmough and Schmoll, 2007). Ditches used to drain wetlands form a permanent surface water connection between isolated wetlands, ditches or streams (Brunet and Westbrook, 2012). The addition of ditches lowers the outlet sills of the depressions, decreasing the volume of water that can be stored. Drainage alters the contributing area through encouraging surface connections, decreasing runoff retention, and more readily allowing these areas to contribute to runoff downstream. Such changes are permanent unless these connections are reversed. Since wetlands naturally retain runoff, the increased surface connectivity due to the drainage of wetlands generally increases the volume of runoff. Up to 71% of the wetland area on the Canadian Prairies was estimated to have been lost to drainage (Environment Canada, 1986; DUC, 2008) by 1986, with an estimated conversion of 1.2 million ha of wetlands to agriculture land from settlement to 1976 (Environment Canada, 1986). Smaller wetlands are more likely to be drained than larger ones (Serran, 2014), and since wetland areas approximate a Pareto distribution

(Shook et al., 2013), the fraction of drained wetlands is much larger than the fraction of wetland area lost to drainage.

The land use of the surrounding basin of a wetland plays an important role in the hydrology of the PPR. It does so by controlling the amount of snow accumulation and re-distribution as well as the amount of infiltration and runoff that occurs (Pomeroy and Gray, 1995; Euliss and Mushet, 1996; Conly and van der Kamp, 2001; Elliott et al., 2001; van der Kamp et al., 2003; Fang and Pomeroy, 2008; Tiessen et al., 2010). The conversion of grassland to cropland has been widespread over the Canadian Prairies since settlement. Both observational (van der Kamp et al., 2003) and modelling (Voldseth et al., 2007) studies have shown that a conversion from grassland to cropland increases the amount of runoff due to a decrease in infiltration. The water levels of wetlands located in agriculturally intensive areas were found to fluctuate greater than the water levels of wetlands located in a more natural grassland setting (Euliss and Mushet, 1996). In the semi-arid region in southern Saskatchewan, one-third of the St. Denis National Wildlife Area was converted from cropland to undisturbed brome grass and resulted in the drying out of the wetlands within the area a few years after the conversion, while wetlands in the cultivated area continued to hold water. The permanent tall grass was able to trap more snow than the cultivated field and increase the amount of infiltration into frozen and unfrozen soils which decreased the amount of water available to runoff (van der Kamp et al., 2003). Granted that brome grass is not native to the Canadian Prairies, Voldseth et al. (2007) identified using model simulations that wetland water levels under brome grass were very similar to native grassland. Given the van der Kamp et al. (2003) study was conducted in the semi-arid region of the Canadian Prairies, the conversion of the upland landcover may not have the same effect in the wetter, sub-humid regions. On the contrary, a modelling study conducted in the Smith Creek Research Basin in southeastern Saskatchewan found that converting the entire watershed into cropland resulted in a decrease in spring surface depression storage due to decreased snow accumulation because of the shorter plant height enhanced blowing snow sublimation (Pomeroy et al., 2010).

Conservation tillage practices, defined as “any tillage system with at least 30% of the residue from the previous crop remaining on the soil surface after seeding” (Tiessen et al., 2010: pg. 964), have been widely adopted across Saskatchewan. Between 1991 and 2011, the amount of seeded land using conservation tillage practices increased from 36% to 90% (Statistics

Canada). Research examining the effects of tillage practices on snow accumulation and runoff has resulted in mixed findings due to locational differences. On the Canadian Prairies, tillage practices were found to have little to no effect on the accumulation of snow, and in turn, snowmelt-induced runoff (Elliott et al., 2001; Tiessen et al., 2010). Elliott et al. (2001) also noted that zero-till initially generated more runoff than conventional tillage, but decreased in time due to an increase in infiltration capability of the soil. Rainfall induced runoff has been found to be less under conservation tillage practices, suggesting that during the cropping season, conservation tillage can effectively reduce runoff during the growing season but may not have an effect during the snowmelt period in cold regions (Tiessen et al., 2010). On the contrary, research has shown that stubble remaining on the fields can act to trap blowing snow (Campbell et al., 1992, Fang et al., 2010) or minimize snow losses to wind (Pomeroy and Gray, 1995). Snow accumulation has been found to be greater in taller vegetation or lower lying areas, like depressions, as snow gets redistributed from hilltops or areas with shorter vegetation due to a greater exposure to the wind (Fang and Pomeroy, 2009).

2.5 Impacts of Changing Climate and Land Use on Streamflow

It is well understood that both climate change and land use changes impact streamflow generation on the Canadian Prairies. Many studies have examined the influence that climate change or land use change, specifically wetland drainage, has on streamflow, while others examined the impacts separately. The following is a summary of what is known from both observational and modelling studies.

In order to isolate the influence of climate change on streamflow across Canada, a study by Burn et al. (2010) looked at climate and streamflow records from the Reference Hydrometric Basin Network (RHBN). The criteria for basin inclusion in this study were the existence of more than 20 years of records with less than 5% of the land surface altered. Due to these stringent criteria, only six stations were available for use on the Canadian Prairies: two were located in south western Saskatchewan and four were located in southern Manitoba. No significant changes to the annual maximum flow magnitude, and only one station in southwestern Saskatchewan showed significant decreasing trends in the timing of the spring maximum flow. In the Canada-wide context, Burn et al. (2010) identified that changes in processes causing maximum flows could switch from snowmelt driven to rainfall driven due to reduced snowfall and increased rainfall portions of total precipitation. This may lead to increased importance of rainfall-runoff flood events, especially if the magnitude or intensity of such events increases.

Fang and Pomeroy (2007) evaluated, through modelling, how future drought conditions could affect snowmelt runoff on the Canadian Prairies. By increasing winter temperatures, and decreasing winter precipitation, fall soil moisture, and vegetation height, the duration of snowcover and amount of snow accumulation were relatively unchanged as the reduced snowfall was compensated by a decrease in the sublimation of blowing snow. Yet, reduced fall soil moisture along with lower precipitation and higher temperatures caused a large reduction in snowmelt runoff, and in turn streamflow. With a 15% decrease in snowfall and a 2.5 °C increase in the winter mean temperatures, it was found that spring streamflow could cease.

Unlike climate change, it is difficult to isolate the changes in streamflow caused by land use changes, specifically drainage. Therefore, most observational studies have analyzed both climate and land use changes when identifying changes in streamflow, and will be discussed shortly. Although few modelling studies have looked at the influence that wetland drainage has on streamflow, the conclusions coincide: drainage increases annual discharge (Gleason et al., 2007; Pomeroy et al., 2010; Yang et al., 2010; Pomeroy et al., 2012) as well as the magnitude and frequency of flooding (Gleason et al., 2007; Yang et al., 2010; Pomeroy et al., 2014). When simulated, the restoration of wetlands to 1968 levels in Broughton’s Creek (Manitoba, Canada) reduced peak flow by 23.4% (Yang et al., 2010). In the Smith Creek Research Basin where wetlands historically covered 24% of the basin area, restoring all wetlands in a model decreased spring streamflow by 79%, whereas draining all wetlands increased spring streamflow by 117% (Pomeroy et al., 2010). In watersheds with less wetland coverage, the effect of draining and restoring wetlands is smaller: restoring all the wetlands in the sub-basins of the Vermillion River watershed (8% historical coverage) reduced annual discharge by ~14%, whereas draining all wetland increased discharge by less than 13% (Pomeroy et al., 2012). It was found that restoring wetlands increases water storage within the basin (Gleason et al., 2007) and reduces the amount of runoff contributing to streamflow, as streamflow generation from runoff is strongly controlled by depressional storage in wetland dominated basins (Shook et al., 2015).

As briefly discussed earlier, the conversion of upland landcover has an effect on the water levels in wetlands (van der Kamp et al., 2003; Voldseth et al., 2007). When examining such conversions at a basin scale, the effects on streamflow slightly differ than those on wetlands. Using a modelling approach, Pomeroy et al. (2010) applied two land cover conversion scenarios to a ~400 km^2 prairie watershed in southeastern Saskatchewan: the first to 100% forest cover and second to primarily agricultural land use, keeping the wetland and open water areas consistent to the 2007 land use. The results from the modelling showed that converting to complete forest cover increased spring discharge by 41% as more snow accumulated due the increased vegetation height which reduced blowing snow sublimation. In turn, the surface depression storage also increased. For complete conversion to primarily agricultural land use, spring streamflow decreased 2% due to an increased loss to blowing snow sublimation due to the shorter height of agricultural plants than forest. Surface depressional storage also decreased due to this loss of snow.

Similar results were found in an observational study across the PPR that looked at meteorological, land use, and streamflow data (Miller and Nudds, 1996). Twelve unregulated rivers from Canada and United States were analyzed for the period 1955 - 1990. Four of the rivers, all within the United States, had significantly increasing trends with no corresponding changes to annual precipitation. Only one basin in the study (Birdtail Creek, MB, Canada) had significantly increasing trends in annual precipitation, yet no corresponding trends in flow rate was identified. The study concluded that landscape alteration, rather than changes in precipitation, had increased runoff into tributaries of the Mississippi River Valley. Based on satellite imagery, locations with greater agricultural modifications of the landscapes were found to have larger annual flow rates. Similar to Burn et al. (2010), the study identified that streamflow on the Canadian Prairies had not changed significantly. Miller and Nudds (1996) suggested that the number of wetlands and extent of untilled vegetation remaining was sufficient to maintain flows, despite alterations from agriculture. The study concluded that wetlands and native vegetation provide natural flood control.

Other studies from around the globe examined potential drivers of change in streamflow and found that not only precipitation but anthropogenic changes influence streamflow. For example, a study in the Atrak River Basin, Iran concluded that changes in precipitation can only partly explain the trends in the hydrological regime observed. Other causes, such as land use change and increased evapotranspiration, are also likely to have played a role in the hydrological regime trends (Sheikh and Bahremand, 2011). Similarly, a study in conducted in the northern Rocky Mountains, Canada analyzed unregulated and naturalized streamflow records for trends after removing the Pacific Decadal Oscillation (PDO) influences (St. Jacques et al., 2010). It was found that declining streamflows were caused by both hydroclimatic and anthropogenic changes, with the later possibly having a greater influence. Current research under the Panta Rhei Scientific Decade (2013 – 2022) of the International Association of Hydrological Sciences (IAHS) is focusing research on improving the understanding of anthropogenic influences on hydrological processes (Montanari et al., 2013).

Two studies examining the long term changes in streamflow on the Canadian Prairies in both regulated and unregulated watersheds concluded that changes in streamflow were primarily driven by changes in precipitation rather than anthropogenic changes such as wetland drainage (Ehsanzadeh et al., 2012a; 2014). In Ehsanzadeh et al. (2012a), streamflow and precipitation records were analyzed in the two river tributaries of the Lake Winnipeg watershed: Red River and Assiniboine River basins. In the Assiniboine River basin, a lack of long-term trends in streamflow and precipitation suggested other gradual changes (such as wetland drainage and farming practices) had not had a detectable effect on streamflow. Precipitation and streamflow were found to have increased in the Red River basin, with streamflow increasing faster than precipitation (300% increase in runoff with a 20% increase in precipitation between 1921 and 2006). It was suggested that the unbalanced change in streamflow was primarily due to an expansion of the contributing area caused by the increased precipitation. Land use changes, such as wetland drainage and agricultural practices, are believed to have little influence on streamflow in the Red River basin due to a lack of effects of land use changes identified in the Assiniboine River basin streamflow records. As a follow up to the identified need in Ehsanzadeh et al. (2012a) for further research to be conducted on smaller sub-watersheds, Ehsanzadeh et al. (2014) examined 17 sub-watersheds across the Canadian Prairies in the Assiniboine River and Saskatchewan River basins. This study concluded that changes in streamflow were attributed to changes in precipitation and other disturbances (such as wetland drainage and ditching) have had little effect on streamflow. The one exception in this study was Smith Creek, Saskatchewan where significant increases in streamflow volume and peak flows between 1975 and 2005 were attributed to anthropogenic causes, such as wetland drainage resulting in extensive loss in wetland area (53%) between 1958 and 2007.

Although it is agreed upon in previous studies that changes in climate effects streamflow (Burn et al., 2010; St. Jacques et al. 2010; Sheikh and Bahremand, 2011; Ehsanzadeh et al., 2012a; Ehsanzadeh et al., 2014), mixed conclusions were found surrounding the effects of anthropogenic changes. Miller and Nudds (1996) concluded that wetland drainage and loss of natural vegetation, not changes in precipitation, caused increases in streamflow in unregulated rivers in the Mississippi River Valley. Many modelling studies have also shown that wetland drainage increases streamflow across various watersheds in the PPR (Gleason et al., 2007; Pomeroy et al., 2010; Yang et al., 2010; Pomeroy et al., 2012; Pomeroy et al., 2014). In contrast, studies by Ehsanzadeh et al., (2012a, 2014) using both regulated and unregulated watersheds concluded that the main driver causing changes in streamflow was from changes in precipitation, not anthropogenic changes such as wetland drainage. The contrasting results in Ehsanzadeh et al. (2012a, 2014) to the rest of the literature may be attributed to a few factors, such as the study period length, use of larger watersheds, as well as the use of both regulated and unregulated watersheds.

2.6 Summary

A 20th century description of the hydrology of the Canadian Prairies is well documented, but the region has recently undergone changes in climate and rapid changes in land use. Recent studies have examined changes in climate and streamflow separately, or integrated them for statistical analysis at the basin scale with mixed results. What is needed is a detailed and comprehensive analysis of the changes in hydrometeorology, land use changes and runoff processes in order to better understand the dimensions of hydrological change in the region, including an analysis of the nuances of these interacting factors that regional studies simply cannot accomplish given their scale.

Chapter 3 : Methods

3.1 Study Site

Smith Creek Research Basin (SCRB) is located in the Assiniboine River Basin, approximately 60 km southeast of the city of Yorkton, Saskatchewan, Canada (Figure 3.1). SCRB has a gross drainage area of 393 km^2 and is relatively flat with slopes of 2-5 % and elevations of 490 - 548 m.a.s.l (Pomeroy et al., 2010). In 2008, the dominant land use was agriculture (primarily cereals and canola; 58%) and the remainder is comprised of native grassland (9%), deciduous woodland (22%) and natural wetlands (11%; Fang et al., 2010). The dominant soil texture is loam (Saskatchewan Soil Survey, 1991). The basin lies within the PPR where large portions of drainage basins do not often contribute runoff to streamflow. The 1:2 year flow non-contributing area has been estimated by Agriculture and Agrifood Canada and is shown along with the prairie agricultural zone and SCRB in Figure 3.1. The non-contributing area calculation is based on data from before the 1980s and since then, wetlands in the basin have been extensively drained. Flows in some wetland drains in SCRB are managed by the Rural Municipalities (RM) of Churchbridge and Langenburg through the use of culvert gates. These are mostly left open but are closed during periods of high spring runoff, resulting in the temporary restoration of certain drained wetlands in high flow years.

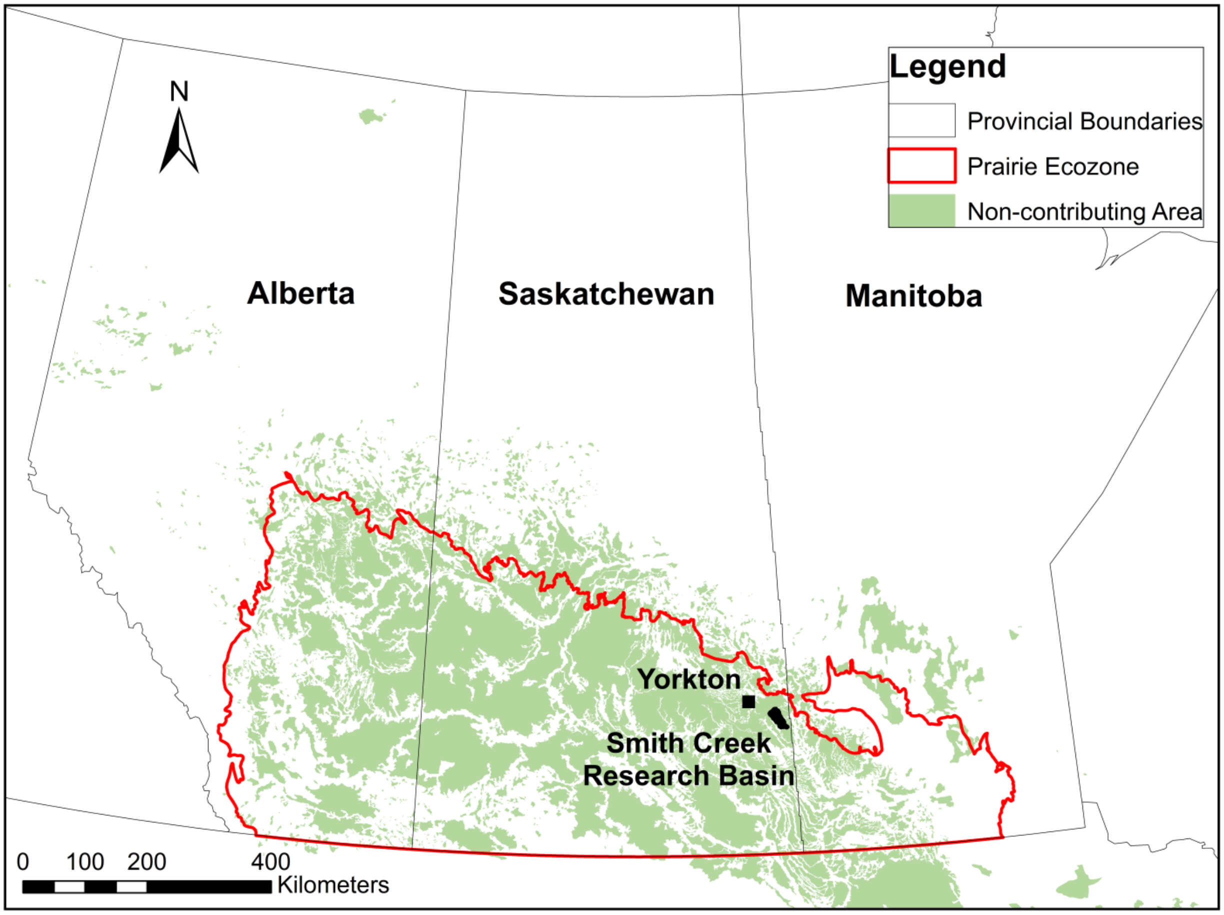

Figure 3.1: Location of Smith Creek Research Basin (SCRB) in the Canadian Prairie agricultural zone (outlined by the red outline). Shaded areas are non-contributing areas in 1:2 year flows as estimated by the Prairie Farm Rehabilitation Administration following Godwin and Martin (1975).

Hydrometeorological instrumentation at SCRB includes a streamflow gauge at the outlet operated by Water Survey of Canada (gauge 05ME007), as well as two meteorological stations nearby each other in the southern part of the basin: one run by the University of Saskatchewan (station name: SC MET) and the other run by Environment Canada (station name: Langenburg, climate ID: 4014145; Figure 3.2). The streamflow gauge has been in operation since, with discharge values being calculated using a rating curve. Daily discharge measurements along with total monthly discharge volumes are available to the public via the Water Survey of Canada website (http://www.ec.gc.ca/rhc-wsc/). The SC MET was installed in July 2007 and measures: relative humidity, air temperature, radiation (incoming and outgoing short and long-wave), wind speed, wind direction, snow depth, soil moisture and temperature (15 cm and 30 cm depth), as well as precipitation using an Alter-shielded Geonor weighing precipitation gauge. The

Langenburg station has been in operation since 1960 and measures temperature and precipitation (rainfall and snowfall) using ruler measurements for snowfall (following a 10:1 relationship, where 10 mm on the ground is equivalent to 1 mm of snowfall) and rainfall was measured using a MSC copper rain gauge and was replaced with a Type B rain gauge in the early 1970's.

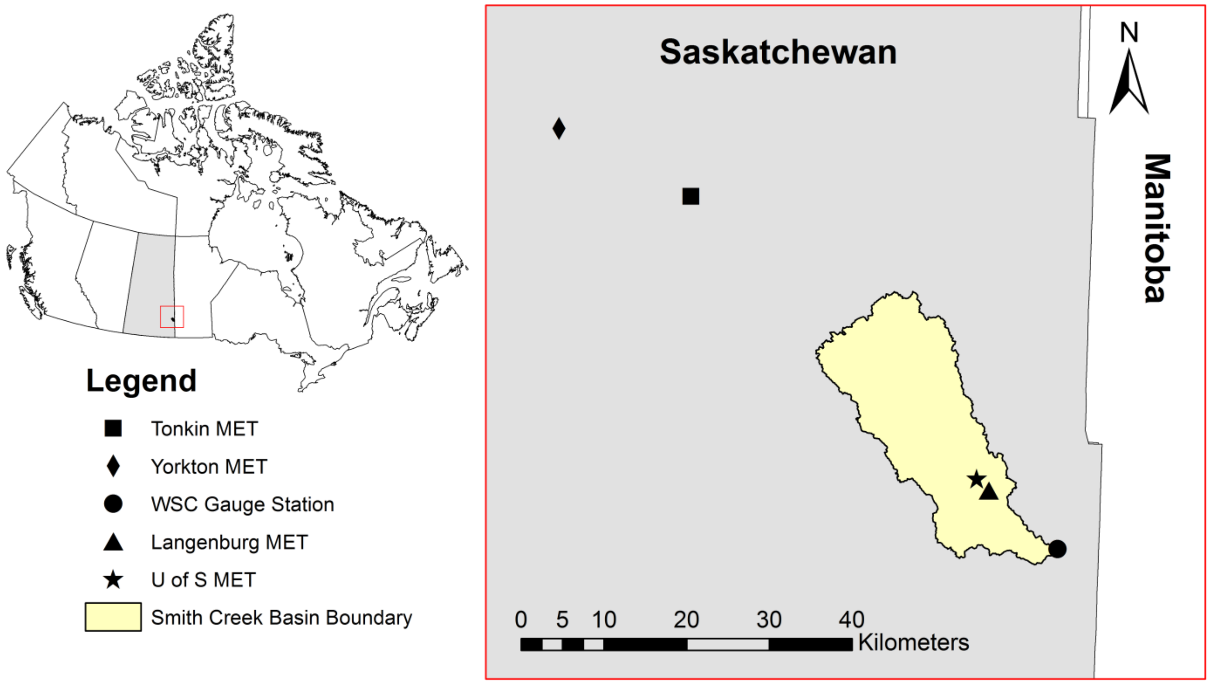

Figure 3.2: Location of all meteorological and gauging stations used in this study.

3.2 Methods

3.2.1 Land Use Change

Summer and fall aerial photographs taken in 1958, 2000, and 2009 by Ducks Unlimited Canada were analyzed to identify changes in ponded depression area and drainage channel lengths (Boychuk et al., 2014). For this study, all open water, marshes and ponds were classified as ponded depressions, as following the classifications set out in the Canadian Wetland Classification System (CWCS; Boychuk et al., 2014). Additionally, the physical delineation of the ponded depression includes the areas that periodically flood, as defined in Gleason et al. (2008; see Figure 3.2 for an example from Boychuk et al., 2014).

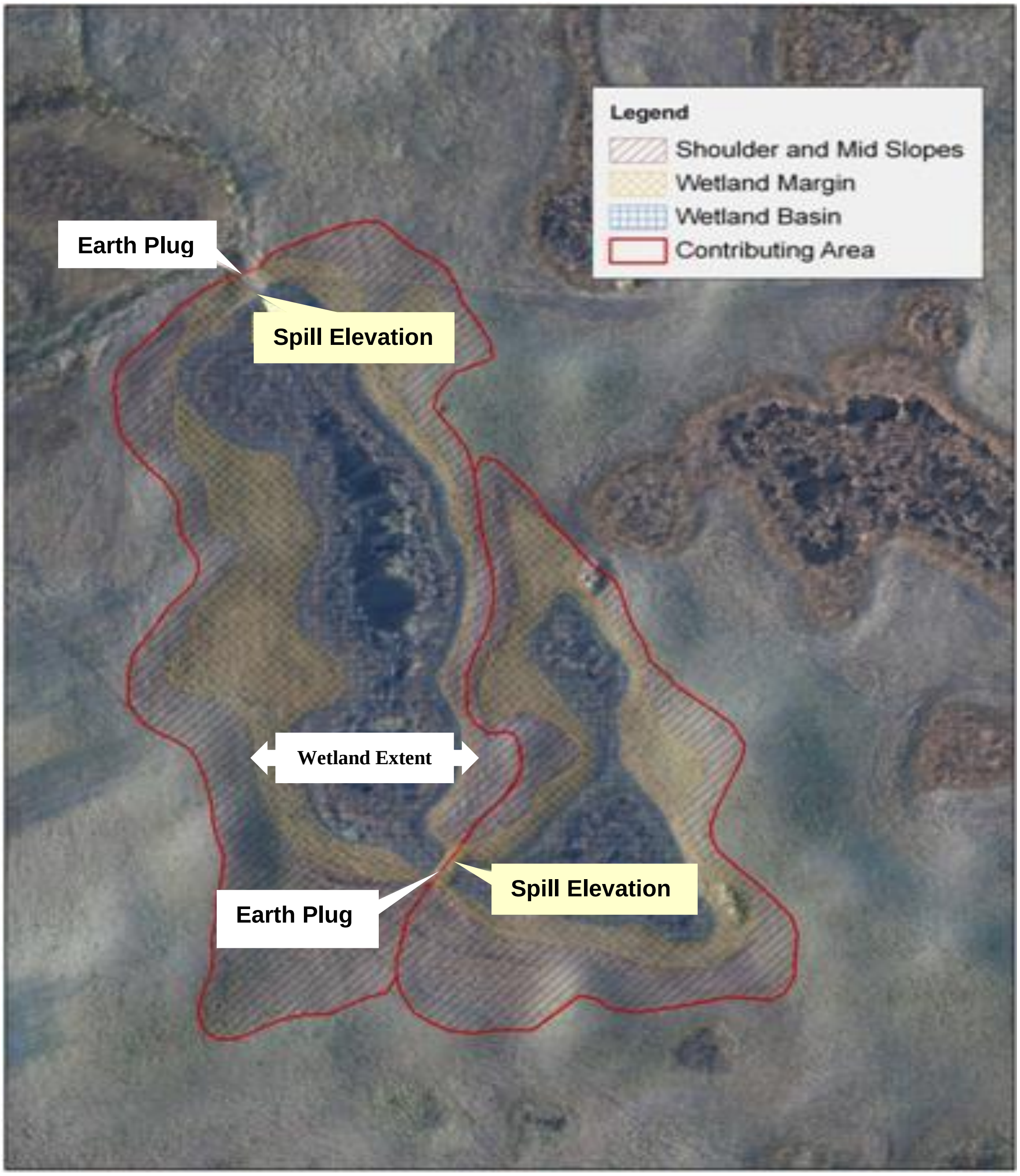

Figure 3.3: An example from Boychuk et al. (2014) showing the delineation of ponded depressions using aerial photographs.

Drainage channels were also identified in areas where changes in ponded depressions were observed. The design of the drainage channels is meant to easily move water off of fields and into the stream network. The description used for identifying an agricultural drainage channel is a "man-made surface ditch with evidence of recent excavation or maintenance" and "represents on-farm drainage" (Boychuk et al., 2014). Three types of agricultural drainage are: internal or terminal ditching, contour ditching, and consolidation ditching. Skeleton vectors were used for connect the inlet and outlet of drains across a ponded depression (Figure 3.3; Boychuk et al., 2014). A more detailed description of the methods used for identifying ponded areas and drainage features can be found in Boychuk et al., (2014).

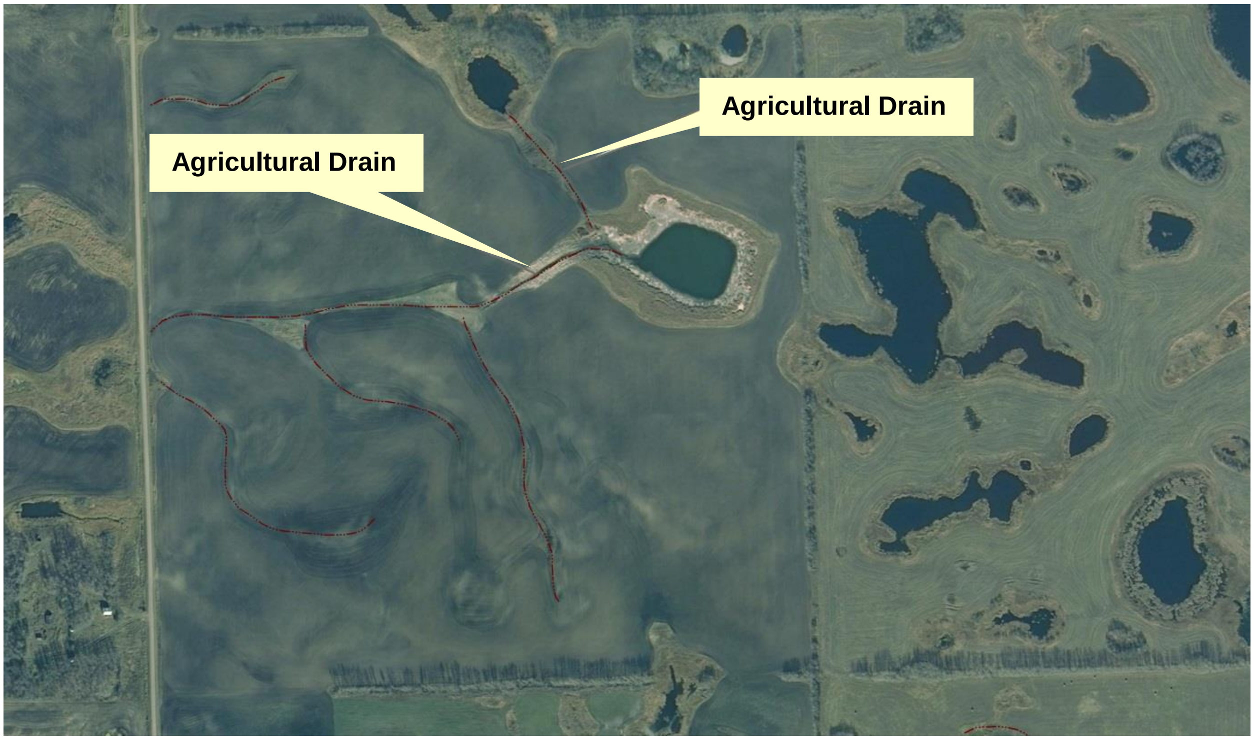

Figure 3.4: An example from Boychuk et al. (2014) showing the delineation of drainage channels using aerial photographs.

The changes in wetland volume were estimated using methods and equations derived in Pomeroy et al. (2009; 2014). LIDAR DEM (digital elevation model) data was available for 2008 and was used to estimate the storage of wetlands using a simplified volume-area-depth method based on work by Hayashi and van der Kamp (2000) and Minke et al. (2010). Although wetland depth cannot be obtained from LIDAR DEM unless the wetland is dry, the changes in depth between multiple area measurements were derived through a simple GIS analysis of the DEM (see Pomeroy et al., 2009 for a more detailed description). Since there was no available LIDAR data for 1958 and 2000, wetland area was used to estimate wetland volume using equations fitted to the 2008 data (Pomeroy et al., 2014). The relationship between wetland volume and area fit a polynomial regression and was used to calculate wetland volume for 1958 and 2000 using wetland area data obtained from air photographs in those years.

Table 3.1: Polynomial regression equations used to estimate maximum wetland storage volumes using wetland areas from aerial photographs from 1958 and 2000. Equations developed using 2008 data.

Sub-basin	Polynomial Regression Equation	R^2
1	$y = 3.77E\text{-}06\ x^2 + 0.548787x - 471.396$	0.996141
2	$y = 3.05E\text{-}06\ x^2 + 0.621025x – 692.822$	0.98279
3	$y = -3.9E\text{-}08\ x^2 + 0.848735x - 1531$	0.990145
4	$y = 4.87E\text{-}06\ x^2 + 0.536234x – 581.972$	0.949545
5	$y = 0\ x^2 + 0.273944x + 7.17845$	0.952344

In order to better represent the varying relationships between wetland area and volume across the SCRB, the entire watershed was divided into five sub-basins. A relationship between wetland area and volume was derived for each sub-basin and can be found in Table 3.1 where *y* is the volume of the wetland and *x* is the area of the wetland in the corresponding year. Further details can be found in Pomeroy et al., (2014).

Historical agricultural land use data was obtained from the Census of Agriculture conducted by Statistics Canada (1961 to 2011). Information for the RM's of Langenburg (No. 181) and Churchbridge (No. 211) were used to document changes in land use for the RM's adjoining and including the SCRB, and involved changes in crop land, summer fallow, pasture land, woodland and wetlands (unimproved land), and tillage practices (only available since 1991).

3.2.2 Precipitation

Daily precipitation (rainfall, snowfall) was compiled for the time period 1942 to 2014. To estimate annual precipitation, daily observations were accumulated from November 1st (of the previous year) to October 31st. This facilitated the association of snow accumulation over the winter months to its melt and flow to the stream each spring. Due to the shortness of meteorological records from the SC MET (2007-2013) and missing daily precipitation data from the Langenburg Station (16% missing; 1960 to present), data from Yorkton, SK (Environment Canada station ID: 4019080; 1942 to present), 60 km to the northwest, was used to both extend the daily precipitation records back to 1942 and infill gaps (see Figure 3.2). Precipitation at the Yorkton station was measured using a copper rain gauge and snow ruler measurements until the early 1970's, then switched to Type B rain gauge and Nipher snow gauge. In 2005, a Belfort precipitation gauge was installed with an Alter shield and replaced in 2011 with a Geonor weighing precipitation gauge.

Precipitation data from the Adjusted and Homogenized Canadian Climate Database (AHCCD) was investigated for use in this study, but short comings were found that made the data inappropriate for use in this study. The AHCCD was developed to produce reliable climate datasets that have been quality controlled, resulting in datasets that have been adjusted for issues such as wind undercatch, wetting losses, and evaporation losses that occur before an observation is made (Mekis and Vincent, 2011). The AHCCD dataset that was examined for this study was Tonkin, SK, which was available from 1942 to 2012. The dataset was compiled from two different Environment Canada stations: Yorkton, SK and Tonkin, SK (located ~20 km east of Yorkton). Yorkton data was used from 1942-2005 and was adjusted by multiplying the values by a 1.2 corrective factor for wind undercatch and snow density. From 2005 and on, data from Tonkin, SK was used in the dataset. It was discovered that there are large differences between the snowfall at Yorkton and Tonkin and the dataset was not adjusted to account for the change (see Figure 3.4). When comparing data from Tonkin and Yorkton, almost twice as much snow was recorded at the Tonkin site, even after the corrective factor of 1.2 was applied to the Yorkton data. From the information on coordinates of meteorological sites, it is likely that the Yorkton site is located in an open area, whereas the Tonkin site may be within a farmer's treed yard. Such differences in station siting explain why Tonkin received more snowfall than the Yorkton site. After discovering this information on how the data was adjusted, it was decided to use Yorkton data to infill gaps in the Langenburg and University of Saskatchewan datasets.

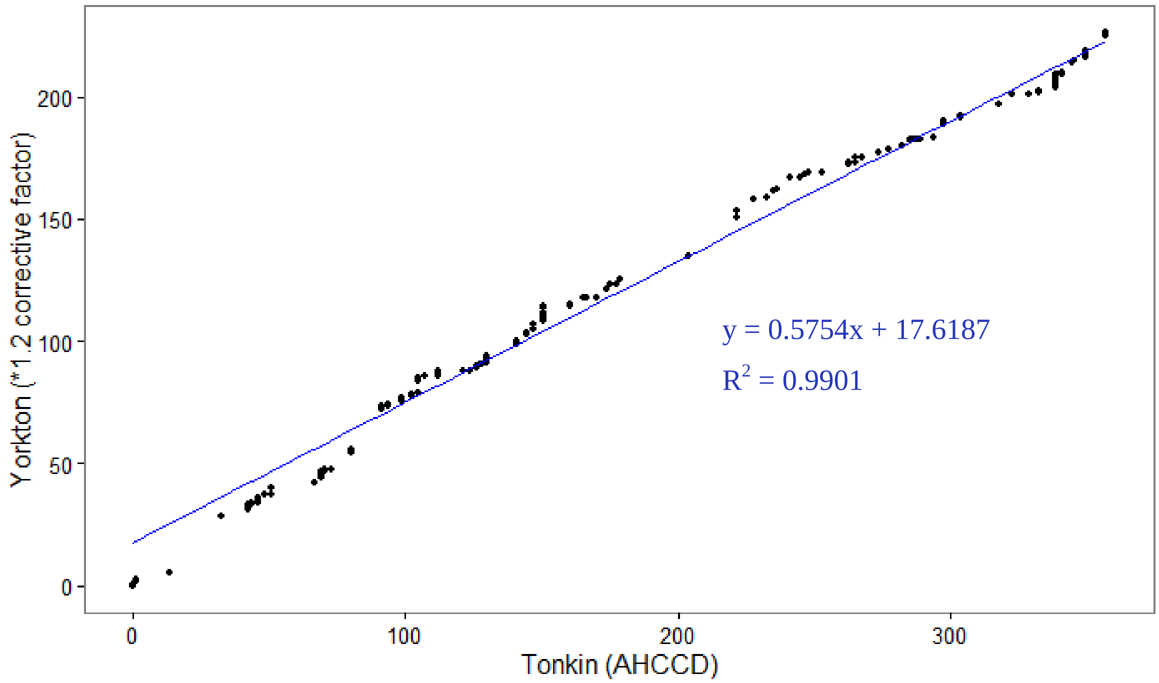

Figure 3.5: Double mass curve of cumulative adjusted Yorkton snowfall data (correction factor of 1.2) and cumulative Tonkin snowfall data from the AHCCD for the period of 2005 to 2007.

3.2.3 Adjusting Precipitation Data

Errors in snowfall measurements can occur due to wind speed, resulting in an undercatch in recorded snowfall (MacDonald and Pomeroy, 2007). As stated in section 3.2.2, AHCCD used a correction factor of 1.2 for snowfall in Yorkton. Although this value is not dependent on wind speed, this value is meant to correct for both undercatch and snow density (Mekis and Vincent, 2011) and so was used to adjust snowfall data from Yorkton for this study. Snowfall data from Langenburg was measured using a ruler and followed a 10:1 rule (10 mm on the ground is equivalent to 1 mm of snowfall). Therefore, the raw Langenburg snowfall data remained uncorrected prior to adjusting for spatial differences between sites (see below).

Snowfall measurements from University of Saskatchewan were adjusted for wind speed on a daily basis using a correction equation derived by MacDonald and Pomeroy (2007) for an Alter-shielded Geonor weighing precipitation gauge:

$$CE_{Geonor} = 1.010\ e^{-0.09\ U} \tag{1}$$

where "CE_{Geonor} is the event based catch efficiency referenced to true snowfall, and U is the wind speed at gauge height (m s^{-1})" (MacDonald and Pomeroy, 2007).

Precipitation data, both rain and snow, from Yorkton and Langenburg differed from the SC MET due to spatial differences. In order to correct for this, data from Yorkton and Langenburg were adjusted using double mass curves to obtain regression equations (see Appendix A). The double mass curve method develops a fixed proportion that is derived from the comparison of two similar cumulative data variables over the same time period (Searcy and Hardison, 1960). These regression equations were then used to adjust the Yorkton and Langenburg data to the SC MET. The cumulative precipitation from the two sites was compared to precipitation from SC MET from 2008 to 2013. Missing data was removed and snowfall data was adjusted (as stated above) prior to comparing sites. Rainfall and snowfall were compared separately as the relationships between sites were different for each precipitation phase.

When comparing the cumulative rainfall and snowfall between Langenburg and SC MET, the slope of the relationship did not vary from 2008 to 2013 (see Appendix A). Therefore, the linear regression equation derived used all available data. The relationship between SC MET and Langenburg snowfall data is relatively weak, and may be due to accuracy of the Langenburg station as snowfall is measured with a ruler. The relationship between the cumulative rainfall and snowfall for Yorkton and SC MET shifted during the 2008 and 2013 time period, signifying changes in relationship caused by changes in method collection or other physical parameters (Searcy and Hardison, 1960). Therefore, regression equations were derived from only partial data, where the slope stayed consistent (see Appendix A for figures). Table 3.2 summarizes the results of the regression equations derived from the double mass curves. Since the regression equations are derived from a comparison of cumulative precipitation variables over the same time period, it has resulted in misleadingly high R^2 values.

Table 3.2: Linear regression results from double mass curves to adjust rainfall and snowfall data for spatial differences.

	Rainfall		**Snowfall**	
	Linear Regression	**R^2**	**Linear Regression**	**R^2**
Langenburg	y = 0.9728x – 16.612	0.9995	y = 0.8857 + 10.656	0.9981
Yorkton	y = 0.9057 + 10.11	0.9979	y = 0.9812x + 1.0811	0.9942

Adjusted Langenburg data was used to infill missing SC MET data, as well as backfill to 1960. Missing Langenburg data was filled in using adjusted Yorkton data, which was also used to backfill to 1942. There were 10 days where all three sites were missing data. Data from Atwater, SK (a site located ~40 km from the study site) was used to infill the gap. The data from Atwater was not adjusted to SC MET as there were no years where both stations were operating at the same time. Figure 3.5 shows the daily precipitation values broken down by site.

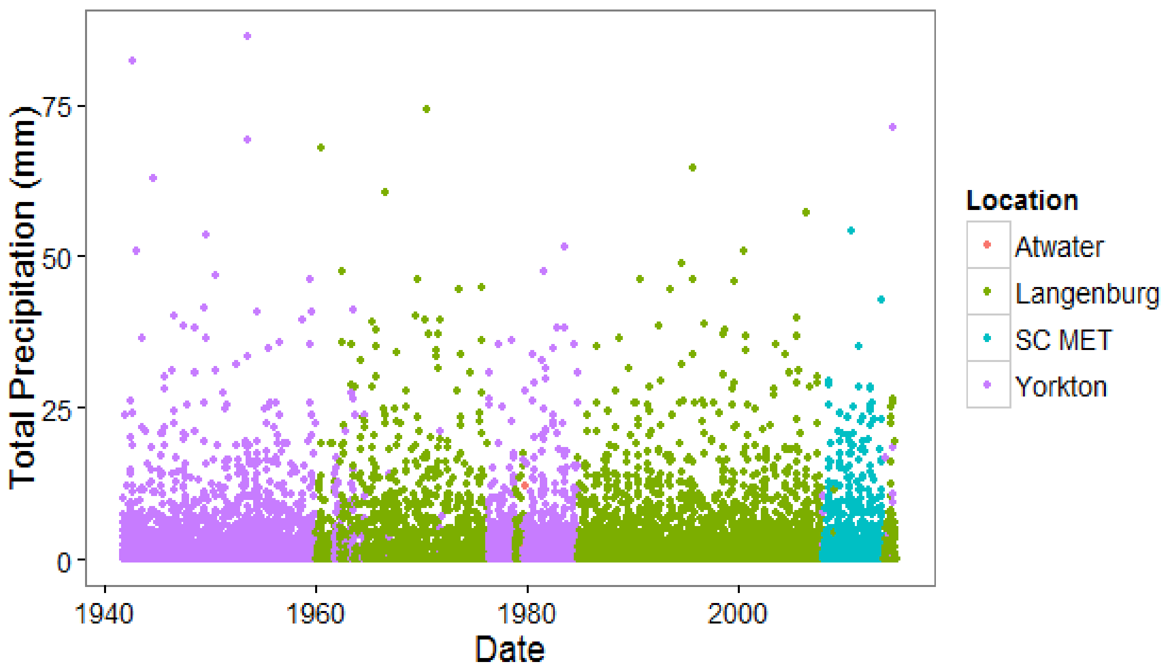

Figure 3.6: Visual depiction of the daily precipitation dataset compiled for this study using adjusted data from four different stations (as described in the methods section).

Trace events are precipitation events of less than 0.2 mm following measurement guidelines set out in the Manual of Surface Weather Observations (MANOBS) as such amounts less are difficult to measure (Environment Canada, 2015). Thus, trace events were not included in this study due to the locational differences of the sites, spatial variability of trace events, and the use of multiple types of precipitation gauges. When comparing the Yorkton, SK precipitation to the AHCCD Tonkin database from 1942 to 2012, trace events added up to 8.5 mm water equivalent (w.e.) of snowfall and 13 mm of rainfall, per year on average. These amounts account for 7% and 4% of the annual snowfall and rainfall, respectively. Together, trace events accounted for less than 5% of the total precipitation on average at the Yorkton site. Precipitation measurements using a Geonor weighing precipitation gauge or ruler measurements are not affected by trace amounts. Therefore, daily precipitation amounts less than 0.2 mm removed from the SC MET records to agree with the precipitation records from Environment Canada stations (Langenburg and Yorkton) which do not include any events less than 0.2 mm, even after a Geonor weighing precipitation gauge was installed in Yorkton in 2011.

Daily rainfall in the SCRB from May to September was separated into single day and multiple day rainfall events. Single day rainfall events are defined as daily rainfall values that are preceded and followed by days with zero rainfall. Single day rainfall events are characteristically convective storms that can be intense with great local depth, but cover small areas and are hydrologically ineffective because the volume of rainfall at a basin scale is small (Shook and Pomeroy, 2012). Multiple day rainfall events have two or more consecutive days where rainfall occurs. These events are typically due to frontal storms, which may have embedded convection, but are generally of lower intensity and cover a much larger area than single day storms. Because of their duration and areal extent, multiple day storms may cause saturation overland flow and streamflow at a basin scale (Shook and Pomeroy, 2012). The time periods before May and after September were not used due to a lack of convective activity as temperatures in spring and fall are usually not high enough to generate convective storms.

Daily snow on ground data was available from Yorkton (1956 – 2012, missing 2007), Langenburg (1994 – 2014) and SC MET (2008 – 2013, missing 2011). Snow on ground data from Yorkton was used in this study due to the long record length. Missing snow on ground data from Yorkton was not infilled using the other stations as snow retention and re-distribution is affected by factors such as vegetation, exposure to the wind, and surface roughness (Pomeroy and Gray, 1995) which vary station to station. From the snow on ground data, multiple variables were extracted and include: maximum snow depth, number of continuous snowcover days, and first snow free date. The maximum snow depth was identified as being the greatest depth of snow, even if it was only for a single day. The number of days with continuous snowcover was gathered by using the longest stretch of time with snow depth greater than or equal to 1 cm. It should be noted that the duration of continuous snowcover does not signify the end of all traces of snow as large accumulations of snow in depressions, channels, or tall vegetation tend to melt at a slower rate and can still dot the landscape after the duration of the continuous snowcover ends. The first snow free date was obtained by looking at the end date of the continuous snowcover days. It is the first date where snow depth is 0 cm. Subsequent snowfall was noted in some years after the first snow free date.

3.2.4 Temperature

Temperature data was obtained from AHCCD Tonkin site from 1942 to 2012 (Figure 3.2). Data for 2013 and 2014 was gathered from Environment Canada Tonkin, SK site (climate ID: 4019082) due to the fact that AHCCD temperature records were available only until 2012. Missing data accounted for ~0.7% and were not infilled. Maximum, minimum, and mean temperatures were analyzed on an annual and monthly time scales using the R Stats Package (R Core Team, 2012) and following a hydrological year (November 1st to October 31st, for annual values). Any year or month with >10% data missing was not included in the analysis.

3.2.5 Streamflow

Streamflow data were obtained from Environment Canada's Water Survey of Canada (WSC) archived hydrometric database online (http://www.wsc.ec.gc.ca). The following briefly describes the gauging station and development of data and a more detailed overview can be found in Appendix C. The gauging station has been operated by the WSC (station 05ME007) since 1975 and is located 3.2 km north of Marchwell, Sask. Figure 3.6 shows a diagram of the gauging station. The gauging station records water levels using a float system in a wood stave stilling well located adjacent to the stream channel with an intake pipe extending into the stream channel. The corrugated metal culvert that passes through the road acts as both a low and high water control.

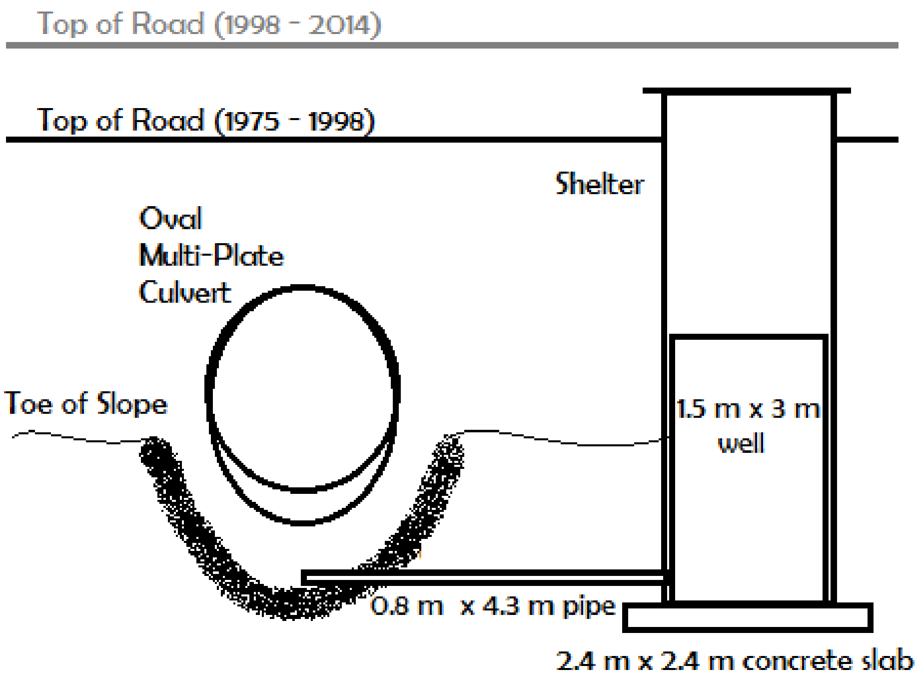

Figure 3.7: Sketch of the Smith Creek gauging station looking downstream (adapted from public records at the Water Survey of Canada).

Manual discharge measurements are derived using the velocity-area method and are obtained approximately 12 times a year by WSC technicians, with an average of 5 of the 12 measurements reading zero flow. The locations of the manual measurements typically occur in the culvert or at a rock wading station 200 m below the gauge, but can vary to other locations depending on the level and safety of wading into the stream. From the manual discharge measurements and recorded streamflow level, a stage-discharge relationship is derived to produce a rating curve which is updated annually. Daily discharge values are produced by the WSC using the rating curves and continuous streamflow levels. Daily discharge values represent the discharge averaged over the entire day, consequently diminishing the instantaneous daily peak value, and in turn, the annual peak discharge. Provided that instantaneous peak flow values were not available for all years, daily discharge values were used for this study.

Uncertainty surrounding peak discharge includes the impoundment of flows due to the numerous culverts and roadways throughout the SCRB. When the flow capacity of the culvert is exceeded, temporary lakes can form upstream of the culvert due to the restricted flow and damming effect of a road above the culvert. Such conditions have been observed at various culvert locations throughout the SCRB, including the gauging station. There are many factors that control the flow capacity of the culvert, including size, shape, length, slope, and roughness of the culvert as well as the inlet design and depth of headwater and tailwater (Smith, 1985), making the flow in each culvert unique. Under high flow conditions when the capacity of the culvert it exceeded, artificial flow through culverts varies from natural, unimpeded flow due as the culverts act to funnel and obstruct streamflow. Therefore, peak discharge values have greater uncertainty during higher flows (peak discharge) than lower discharges because the culvert at the gauging station was inundated during peak flow in 1995, 2011 and 2014. Many culvert improvements have been made throughout the SCRB and range from increasing the culvert sizes to adding in a second culvert. At the gauging station located at the outlet of the SCRB, improvements were made in September 1998 in response to the flood of 1995 when streamflow threatened to wash out the roadway. To reduce the potential for future washouts, the roadway going over the culvert was built up higher, although it is unknown exactly how much higher. During this construction, the culvert was replaced to account for the increased road height, requiring the culvert to be longer in length due to the increased based width of the road. The diameter of the culvert remained unchanged during this process. Other culvert and roadway changes made throughout the SCRB are not available.

Streamflow data from 2014 is provisional as the corrected streamflow data from the WSC is not available until summer of 2015. Streamflow data for 2014 was gathered from the real-time data available on the WSC website (https://wateroffice.ec.gc.ca/) which is available at 5 minutes time intervals. From the real-time 5-minute data, daily discharge was calculated using the average discharge over a single day. The Smith Creek gauging station was inundated in 2014 and failed on June 30th during the rising limb of the flood event. This resulted in 22 days of missing data. A manual measurement was taken on July 1st, and is considered to be the approximate peak discharge of the event. Missing data also occurred for 18 days in late August and early September. All missing data was linearly interpolated between manually measured and station measured points for the purpose of this study.

Using the daily discharge data for Smith Creek from the WSC, annual and snowmelt peak daily flowrate was compiled. The annual daily peak flowrate is identified as the largest daily flow value in a given year (March to October), whereas snowmelt peak daily flowrate represents the peak daily streamflow derived from snowmelt runoff (March to May). The corresponding dates to the peak flowrates were also gathered. In order to accurately determine the timing and magnitude of spring peak flows, the plots of the daily flow values were manually viewed. Daily temperature, precipitation and snow on ground data were also used to aid in the determination of the snowmelt peak. Both the annual and snowmelt peak daily flowrates were examined as snowmelt peak daily flowrate was not always the peak flowrate for that given year.

Runoff ratios were calculated based on the annual depth of runoff and annual precipitation to represent the runoff response of the landscape. The depth of runoff was calculated by dividing the total volume of runoff by the basin area. To calculate the runoff ratio, the depth of runoff (in mm) was divided by the annual precipitation (mm). The result is a ratio of total precipitation that which runs off and contributes to streamflow at the outlet of the basin.

Streamflow was categorized into various runoff mechanisms: snowmelt, rainfall, and mixed snowmelt and rainfall. Hydrograph separation techniques are typically used to separate out the base flow portion of streamflow or other source components by using graphical separation or natural occurring tracers, such as isotopes (McNamara et al., 2007). Streamflow in Smith Creek was assumed to have little to no deep groundwater contributions as the stream typically dries out after snowmelt runoff is complete. Therefore, typical hydrograph separation techniques could not be used for this study. Instead, a set of rules were developed for this research to categorize streamflow into various runoff mechanisms using observed daily temperature, rainfall, snowfall, discharge and snow on ground data.

Streamflow was classified as being derived from snowmelt at the beginning of the spring freshet, which corresponds to snowmelt. It remained classified as snowmelt until a daily rainfall of > 5 mm occurred concomitantly with increasing discharge. If the discharge was >0.1 m^3/s prior to the rainfall event or snow depth was >0 cm, then the streamflow was classified as a mixed snowmelt and rainfall runoff regime. If such conditions occurred on the rising limb of the hydrograph, the streamflow was classified as mixed snowmelt and rainfall for the duration of the rainfall event and returned to a snowmelt runoff regime afterwards. If such conditions occur on the falling limb of the hydrograph, streamflow is classified as mixed snowmelt and rainfall until streamflow <0.1 m^3/s. If discharge is <0.1 m^3/s prior to the rainfall event and there is no recorded snow on the ground, the streamflow is classified as a rainfall runoff regime (see Figure 3.7 for an example).

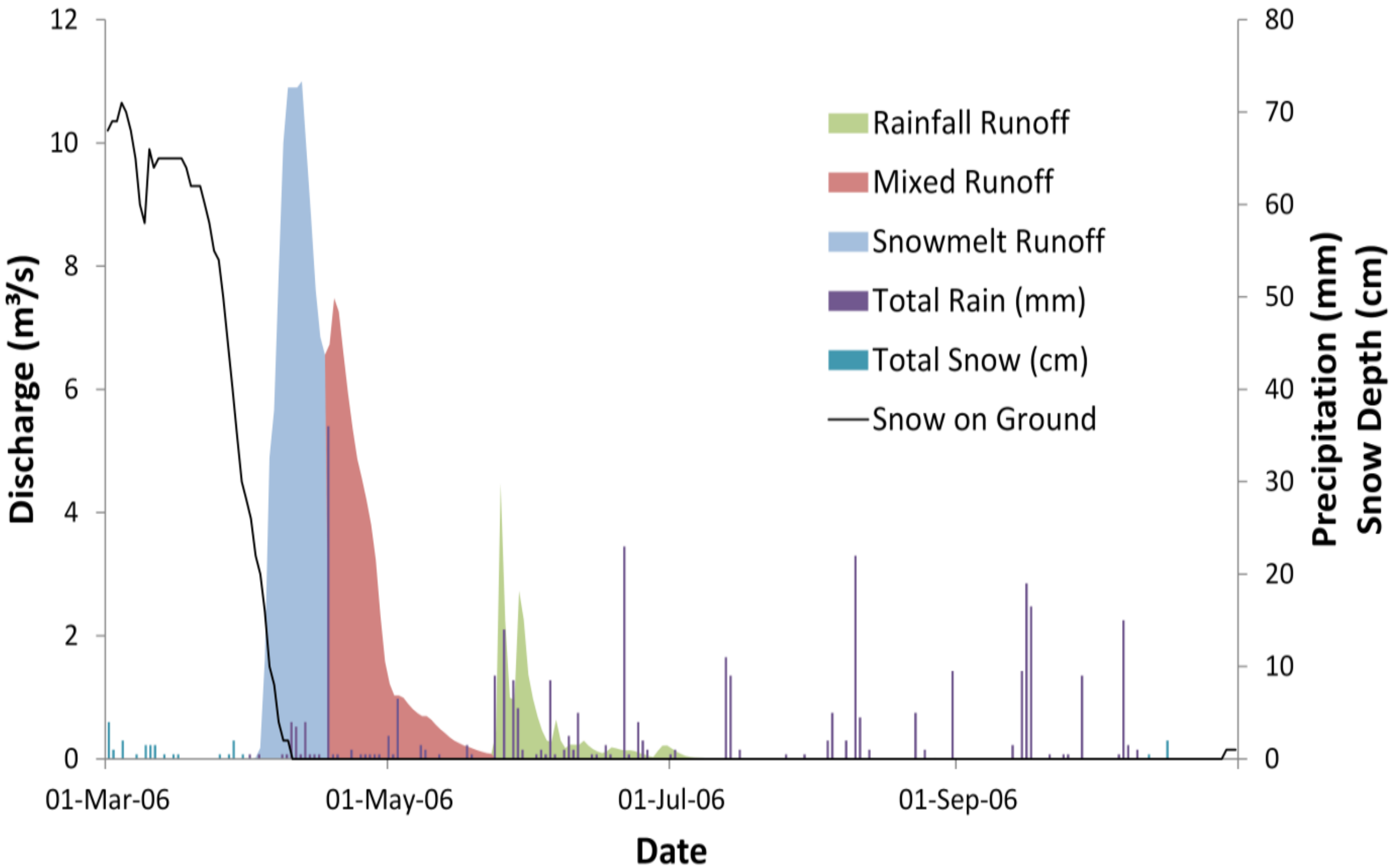

Figure 3.8: An example showing the streamflow separation method which categorizes streamflow into snowmelt, rainfall and mixed snowmelt & rainfall runoff mechanism using observed meteorological data for 2006 at Smith Creek, SK.

A rainfall threshold of 5 mm was used in order to exceed daily evaporative losses and have the ability to be hydrologically effective, meaning runoff derived from the rainfall event reaches the stream channels. Modelling results in Armstrong et al. (2015) identified that in the sub-humid zone of the Canadian Prairies, early spring evaporative losses from saturated bare soils ranged between 5 to 8 mm/day and may represent unrestrictive evaporation rates prior to the growing season. Since the rules state that discharge concurrently needs to increase with a rainfall greater than 5 mm, it omits the rainfall events >5 mm that did not have an influence on streamflow. The discharge threshold of 0.1 m^3/s was determined by examining the daily streamflow data. The average increase in streamflow due to a rainfall event > 5 mm was 0.13 m^3/s and ranged from 0.01 m^3/s (occurred when streamflow resumed in summer after a rainfall event) to 10.36 m^3/s (in 2014 when streamflow responded to a large rainfall event). For streamflow to be classified as dominated by rainfall runoff processes, discharge prior to the rainfall event needed to be less than the average discharge increase during an event (0.128 m^3/s). Therefore, a threshold of 0.1 m^3/s prior to the rainfall event was used to allow for streamflow to transition to being rainfall runoff dominated.

There is uncertainty surrounding the categorization of streamflow events attributed to the various runoff mechanisms due to the abrupt transitions from one mechanism to the other as well as the threshold values chosen. In reality, streamflow does not abruptly transition from one type of runoff event to another. But for the purpose of this study, the lack of natural tracer information and use of daily observed meteorological and streamflow data results in difficulties identifying the fraction of daily streamflow derived from the individual runoff mechanisms. This type of abrupt change in runoff mechanism used for this study results in larger uncertainty in wetter years when it takes longer for streamflow to recede to the low discharge threshold. The information gained when looking at longer time scales (multiple years or decades) can be useful in understanding the general trends and changes in runoff processes that may have occurred, with the greatest uncertainty surrounding the annual volumes of streamflow derived from each runoff mechanism.

3.3 Statistical Analysis

3.3.1 Trend Tests

All observed variables were examined for the existence of trends at daily, monthly and annual time scales. The nonparametric Mann-Kendall (MK) statistical test was chosen for its ability to handle non-normally distributed and missing data, as are commonly in hydrological datasets (Hirsh and Slack, 1984).

The following explanation of the MK test is taken from Yue et al. (2002) and Sheikh and Bahremand (2011). The null hypothesis H_0 states that the data ($X_1, X_2, \ldots, X_n$) are independently and identically distributed random variables. The alternative hypothesis H_1 is that a monotonic trend exists in the data (X). The MK test is based on the test statistic, S, which is defined as:

$$S = \sum_{i=1}^{n-1} \sum_{j=i+1}^{n} sgn(X_j - X_i) \tag{2}$$

where X_j and X_i are the sequential data, n is the length of the dataset, and

$$sgn(\theta) = \begin{cases} = 1 & if\ \theta > 0 \\ = 0 & if\ \theta = 0 \\ = -1 & if\ \theta < 0 \end{cases} \tag{3}$$

A positive S signifies an upward trend whereas a negative S signifies a downward trend. The test statistic S has a mean of zero and the variance is calculated as:

$$Var(S) = \frac{n(n-1)(2n+5) - \sum_{i=1}^{m} t_i(i)(i-1)(2i+5)}{18} \tag{4}$$

where m is the number of tied groups (equal observations), each with t_i tied observations (i.e. number of data in the tied group).

The standardized test statistic Z follows the standard normal distribution with a mean of zerio and variance of 1. It is calculated as:

$$Z = \begin{cases} \frac{S-1}{\sqrt{Var(S)}} & S > 0 \\ 0 & S = 0 \\ \frac{S+1}{\sqrt{Var(S)}} & S < 0 \end{cases} \tag{5}$$

If $|Z| < Z_{1-\alpha/2}$, the H_0 should be accepted.

All variables were tested for serial correlation at 95% confidence levels prior to the MK trend test as serial correlation can influence test results (Yue et al., 2002). Both of the Mann-Kendall and the ACF tests were conducted in R using the 'Kendall' (McLeod, 2011) and 'R Stats' (R Core Team, 2012) packages. Trends were considered significant at $p < 0.05$. Temporal trends in continuous data were identified using linear regression (Ebdon, 2004), whereas logistic regression was used on discrete data (Hosmer et al., 2013) using the 'trend' package (Frei, 2013).

3.3.2 Changepoint Analysis

In order to identify shifts or changes in the direction of trends, change point analysis was conducted in R using the 'changepoint' package (Killick and Eckley, 2013). In this package, the Segment Neighbourhood identifies changepoints in the mean and variance. According to Auger and Lawrence (1989), "a segment neighborhood is a set of contiguous residuals that share common features". The following is a description of the segment neighborhood algorithm from Auger and Lawrence (1989) and Killick et al. (2012). See references for further detail.

Segment neighborhoods are identified by optimally partitioning a sequence into Q contiguous segments based on the fit of the model to the data. It does so by searching the entire dataset using dynamic programming and starts with setting an upper limit on the maximum number of changepoints (Q). The cost function for all possible segments are computed and all possible segmentations between 0 and Q changepoints are considered. Let where Y_i = vector of observed data of i^{th} residue; θ_q = vector of unknown parameters of q^{th} segment neighborhood; r_q = unknown index of last residue of the q^{th} segment neighborhood (segment neighborhood boundaries); $F(Y_i \ldots Y_j, \theta_q)$ = model of the relationship between $Y_i \ldots Y_j$ and θ_q (model). To find estimators for the segment neighborhood parameters ($\theta_1,\ldots,\theta_Q$) and boundaries ($r_1, \ldots, r_{Q-1}$), the approach is to minimize:

$$Z(Y,\theta,r,Q) = \min_{\theta,r} \sum_{q=1}^{Q} C(F\left(Y_{r_{q-1}+1} \ldots Y_{r_q}, \theta_q\right)) \tag{6}$$

Where C = function that measures the fit of the model $F()$ to the data $Y_{rq-1+1} \ldots Y_{rq}$, with the estimated parameters $\theta_1,\ldots,\theta_Q$ (objective function) and $(r_0, r_Q)=(0,n)$ where n is the length of the sequence.

The maximum number of changepoints is either set by the algorithm, or can be specified by the user. This proved useful when dealing with the datasets as some changepoints detected were short when no manual maximum was specified (1-3 years in length). Changepoints that were three years or shorter were deemed to be spurious, therefore, if results contained any changepoints less three years in length, the maximum amount of changepoints were reduced by one unit until all changepoint lengths were greater than three years.

3.3.3 Teleconnections

Climate teleconnections such as the El Niño Southern Oscillation (ENSO), Pacific Decadal Oscillation (PDO) and others have been shown to influence some annual streamflows in the Canadian Prairies; the influence is more frequently observed in the western portion of the region (St. Jacques et al., 2014). The influence of teleconnections on streamflow was conducted following the methodology of St. Jacques et al. (2010) and Harder et al. (2015) which uses Generalized Least Squares regression to model the impacts of climate oscillations. For this study, the average annual streamflow (based on daily flow values) and annual peak streamflow were examined for the influence of the Southern Oscillation Index (SOI) and PDO. The PDO index was obtained online at http://research.jisao.washington.edu/pdo/PDO.latest (accessed January 27, 2015). The winter (November to March) monthly PDO index value was averaged and used to compare to the following streamflow season (March to October). The standardized SOI data was obtained from http://www.cpc.ncep.noaa.gov/data/indices/soi (accessed January 27, 2015). The previous summer and fall (June to November) SOI values were compared to the following streamflow season. R was used (R Core Team, 2012) with code by Harder et al. (2015).

Chapter 4 : Results

4.1 Land Use

From 1958 to 2009, the SCRB has been partially drained. In 1958, ponded area accounted for 24% of the basin area and it dropped to 12% in 2000. Further drainage resulted in ponded area accounting for 10% of the basin in 2009 (Figure 4.1, Table 4.1). The length of drainage channels has also increased substantially since 1958 (Table 4.1). A four-fold increase was observed between 1958 and 2000, whereas an eight-fold increase was observed from 1958 to 2009. The difference between the two time periods suggests that the rate of drainage in the 21st century has increased. Local farmers report that efforts to drain wetlands are increased following wetter spring conditions.

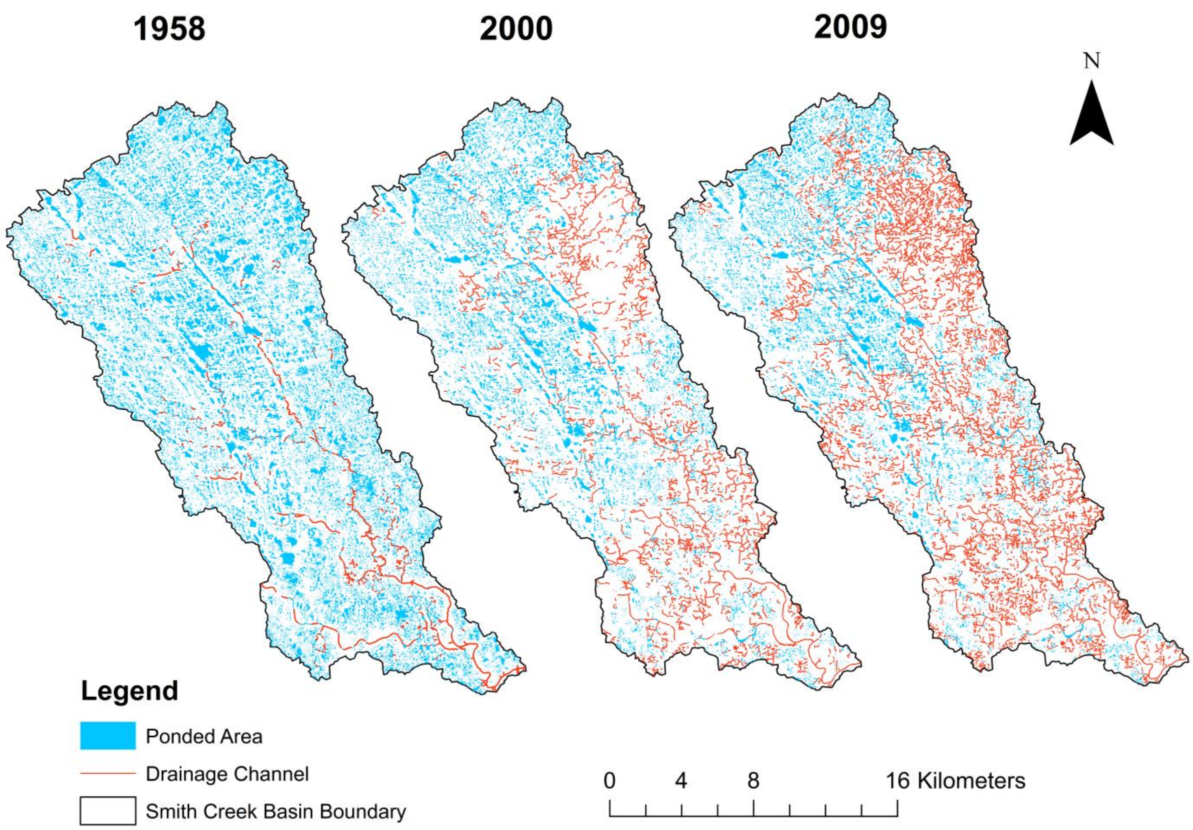

Figure 4.1: Ponded area and drainage network in SCRB in 1958, 2000 and 2009. Data provided by Lyle Boychuk, Ducks Unlimited Canada from aerial photograph analysis and mapped for the basin area determined by Fang et al. (2010).

Table 4.1: Changes in ponded area and drainage channel length measured by aerial photograph analysis from 1958 to 2009 and changes in maximum wetland storage volume from 1958 to 2008 in SCRB.

	1958	2000	2008/2009
Ponded Area (km^2)*	96.0	47.0	40.3
Ponded Area (%)*	24	12	10
Drainage Channel Length (m)*	119,348	503,722	931,312
Maximum Wetland Storage Volume (m^3) †	3,568,182	957,364	766,325

*: Results from Boychuk et al. (2014)
†: Results using equations in Table 3.1

Maximum wetland storage volume estimates using equations in Table 3.1 show a considerable decline between 1958 and 2008. From 1958 to 2000, wetland storage volumes declined 73%, and from 2000 to 2008, they declined a further 20%. Overall, there has been a 79% reduction in the maximum wetland storage volumes, resulting in only 1/5th of the original volume remaining by 2008.

Substantial changes in agricultural land use and tillage practices have occurred between 1961 (Statistics Canada, 1961) and 2011 (Statistics Canada[1]). Crop land has been the dominant agricultural land use, increasing from 33% of total farm area to 60%. Pasture land increased from 2% to 9%, while summer fallow decreased from 20% to 5% of total farm area. Woodland and wetlands (unimproved land) decreased substantially from 46% to 27%. Between 1991 and 2011, the adoption of zero till practices has been considerable, increasing from less than 2% of seeded land to 34%. Conservation tillage (tillage retaining most of the residue on the surface) also increased from 25% to 47% whereas conventional tillage (tillage incorporating most crop residue into the soil) decreased substantially from 73% of seeded land to 19%.

4.2 Temperature

Changes in meteorological variables observed in SCRB signify a warming trend. The average annual daily maximum, minimum and mean temperatures in SCRB for 1942 to 2014 are 7.3, -3.9 and 1.7°C, respectively. The annual values are averaged over a hydrological year spanning from November 1st to October 31st. Annual daily maximum temperature has increased by 1.2 °C ($p = 0.030$), with no significant changes to mean and minimum annual temperatures (Figure 4.2, Table 4.2). On a monthly scale, significant increases in temperatures were found for

January, February, March, June, and September and ranged from 1.1 to 4.0 °C. March temperatures increased the most with an increase of mean monthly temperature of 3.5°C. In contrast, October mean and minimum temperatures have slightly decreased.

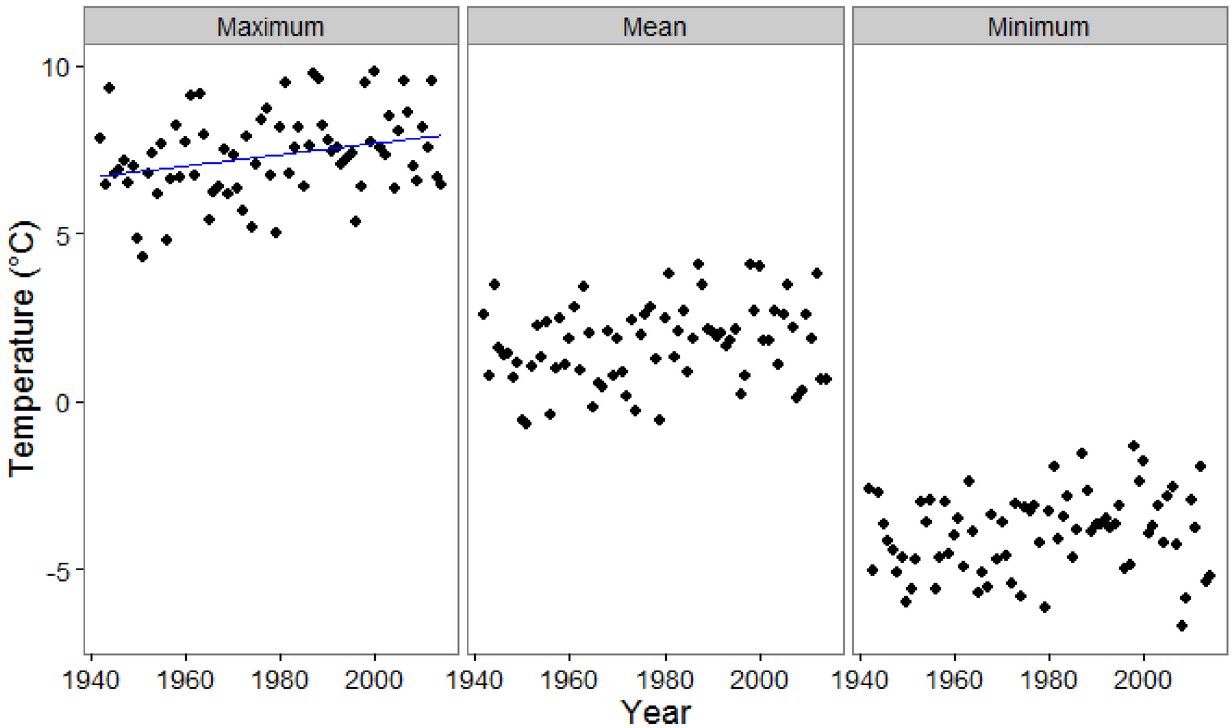

Figure 4.2: Annual maximum, mean, and minimum temperatures following a hydrological year of November 1st to October 31st from Tonkin, SK (1942 – 2014). Blue line represents significant linear trend at 5 %.

Table 4.2: Changes in annual and monthly temperature (°C) with significance determined via a Mann Kendall test ($p < 0.05$).

	Maximum	**Minimum**	**Mean**
Annual	+ 1.2	–	–
January	+4.0	–	–
March	+3.4	+3.5	+3.5
June	+1.2	+1.1	+1.1
September	+3.2*	–	+2.1*
October	–	-1.4	–

* *p<0.01*

4.3 Precipitation

Mean annual precipitation from 1942 to 2014 was 442 mm, with rainfall accounting for 325 mm (73%) and snowfall 117 mm (27%; Figure 4.3). Although there are no significant trends ($p > 0.1$) or shifts in total annual precipitation, the increasing temperatures coincide with a gradual change in precipitation phase. Annual rainfall has significantly increased ($p = 0.013$) at a rate of 0.9 mm/year ($p = 0.013$) while annual snowfall has insignificantly decreased by 0.5 mm/year ($p = 0.102$). The result is a significant increase in the annual rainfall fraction of precipitation ($p = 0.043$; Figure 4.4) from 68% to 78% of annual precipitation between 1942 and 2014. Concurrently, snowfall fractions gradually decreased from 32% to 22%.

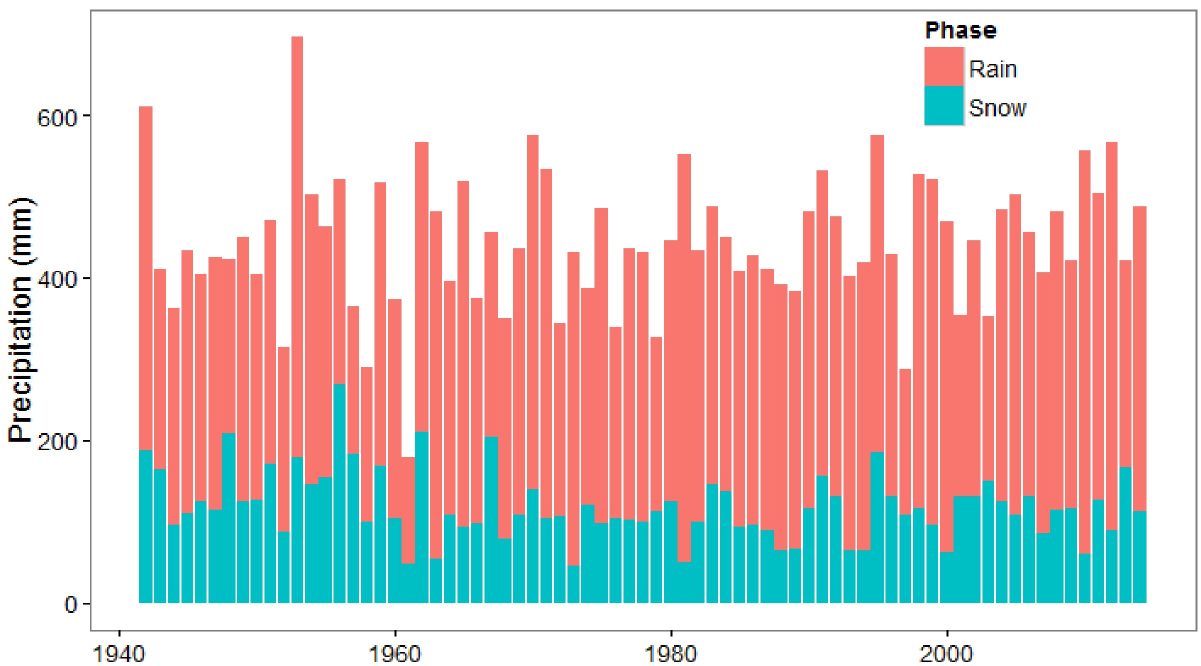

Figure 4.3: Annual rainfall and snowfall at Smith Creek, SK from 1942 to 2014 following a hydrological year of November 1st to October 31st.

On a monthly scale, rainfall fractions have significantly increased in March ($p = 0.006$) with a changepoints in 1963 and 1972, that ultimately quadrupled the March mean rainfall fractions from 4.8% to 19% (Figure 4.4). Monthly rainfall fractions were only found to be changing in March, but there were also significant increases in the amount of rainfall in March ($p = 0.010$), May ($p = 0.034$), and June ($p = 0.048$). Rainfall amounts in October also increased, yet the significance level did not quite meet the 5% significance level ($p = 0.051$). Mean monthly rainfall gradually shifted to the early summer months between 1942 and 2014, increasing from

31 to 59 mm in May and 58 to 92 mm in June. There was also a shift in phase from snowfall to rainfall in March and October with rainfall increases from 1 to 4 mm in March and 14 to 23 mm in October (Figure 4.5). No other months were found to have significant increasing or decreasing trends in monthly rainfall, snowfall, or total (rainfall + snowfall) precipitation.

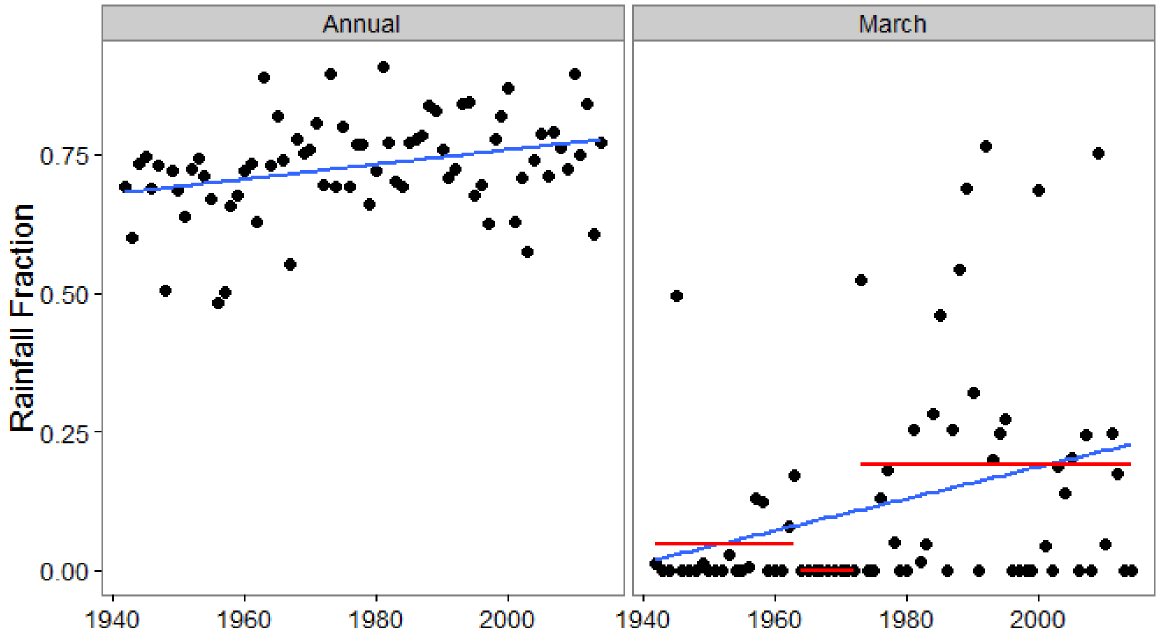

Figure 4.4: Rainfall fractions on an annual basis and for the month of March, both of which are significantly increasing. For March, changepoints were identified in 1963 and 1972, with the mean over quintupling over the time period. Significant linear trends ($p<0.05$) are represented as blue lines, changepoints are represented as red lines.

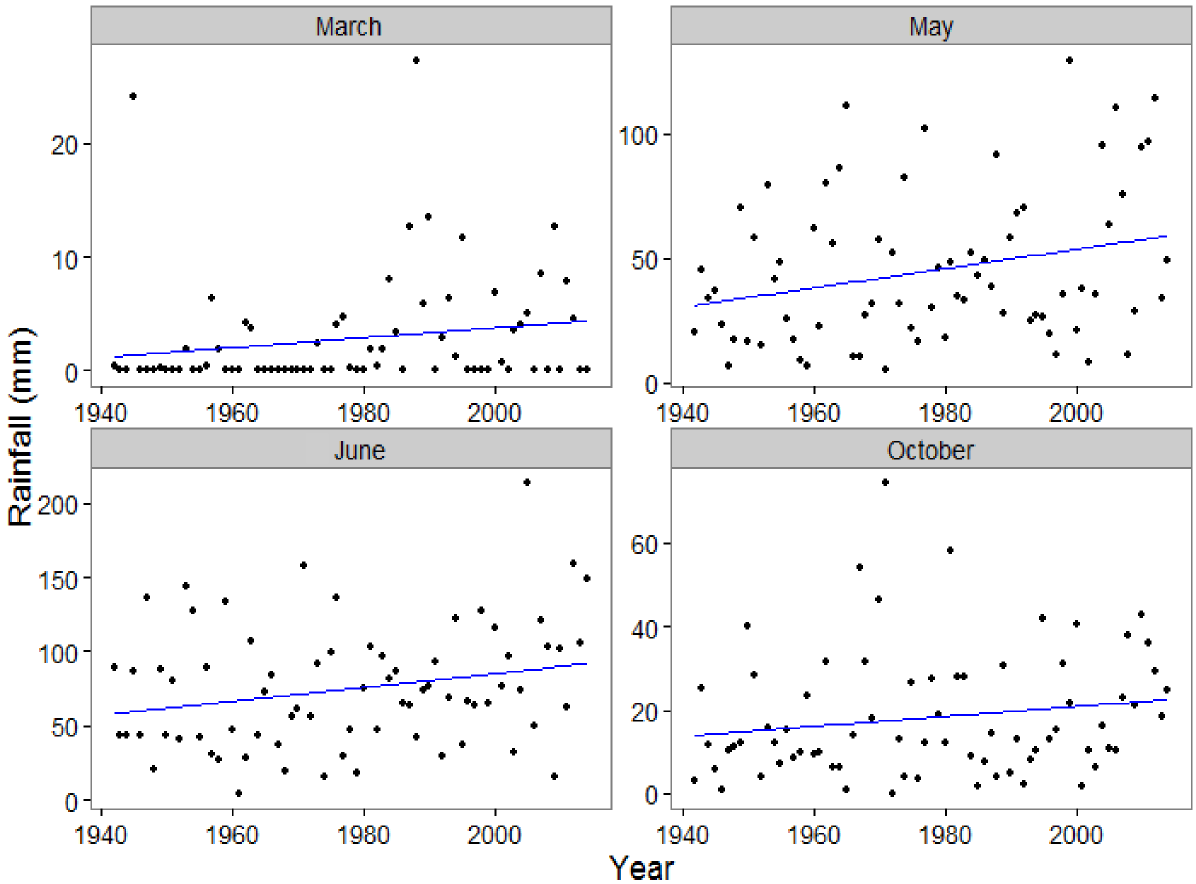

Figure 4.5: Monthly rainfall amounts for March, May, June, and October 1942 to 2014. Note: y-axis scales vary, only statistically significant trends shown at p<0.05 and represented by the blue line.

The duration of rainfall events during the summer months has also changed throughout the study period (May to September). Single day rainfall events are characteristically convective, single day storms whereas multiple day rainfall events have two or more consecutive days where rainfall occurs and are typically frontal events (Shook and Pomeroy, 2012). The analysis was restricted to the period of May to September as that is when single day convective rainfall events can occur. The number of multiple day events has increased significantly by 50% (p = 0.029, Figure 4.6), yet the number of single day events have not changed significantly (p > 0.1), resulting in an increase in the total number of rainfall events of 14% (p = 0.025). From 1942 to 2014, the annual number of summer rainfall events has increased by 3.5, whereas multiple day summer rainfall events have increased by 5. The increase in multiple day rainfall events is mainly due to an increase in the number of 2-day events, accounting for 3 of the 5 increased multiple day events (p=0.004).

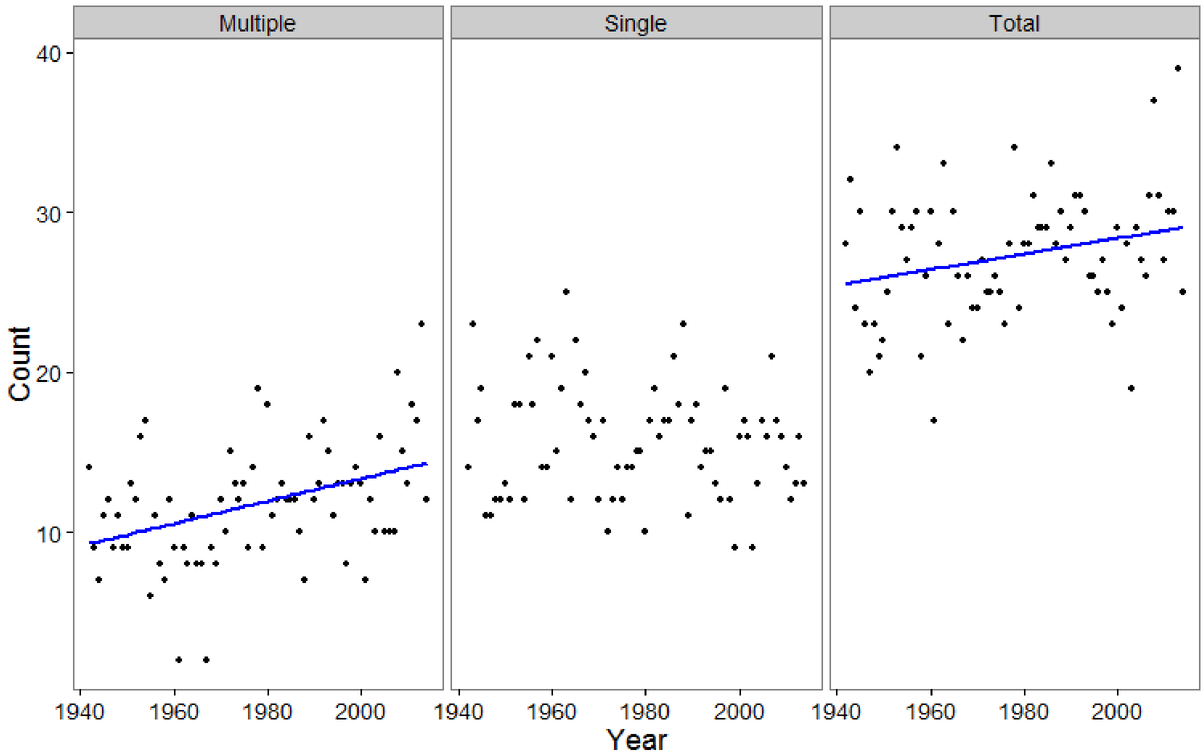

Figure 4.6: Total number of summer rainfall events split up into multiple day, single day, and total rain days from 1942-2014. Blue line represents statistically significant linear trend at 5% significance.

The precipitation regime on the Canadian Prairies is dynamic in nature, resulting in wet and dry years. In order to normalize the data to the number of events per year, the fraction of multiple day and single day summer (May – September) rainfall events were calculated on an annual basis. The fraction of multiple day events is the number of multiple day events divided by the total number of events for the given summer. Results can be seen in Figure 4.7. Single day and multiple day rainfall event fractions account for 0.58 and 0.42 of the total annual summer rainfall events on average from 1942 to 2014, respectively. It was found that the fraction of rainfall events falling as multiple day events has increased significantly from 0.37 to 0.48, while the fraction of single day rainfall events has decreased from 0.63 to 0.52 ($p = 0.013$) of the total summer rainfall events. This means that despite varying wet and dry years, a higher fraction of rainfall events in the summer months are falling as multiple day events which may be associated with frontal precipitation systems which have the ability to generate rainfall runoff on a basin scale compared to single day events which are mainly due to convective storms of small spatial scale. Although the increase in rainfall in the summer months is predominately in the form of multiple day events, the magnitude of all individual rainfall events has not increased significantly (categorized into all single or all multiple day events from 1942 to 2014; $p > 0.1$). When categorizing the individual multiple day rainfall events into the duration of the event (2-day, 3-day, 4-day, and 5+ days), there is some evidence that the magnitude of the 2-day events has increased by 2 mm per event since 1942 ($p = 0.094$).

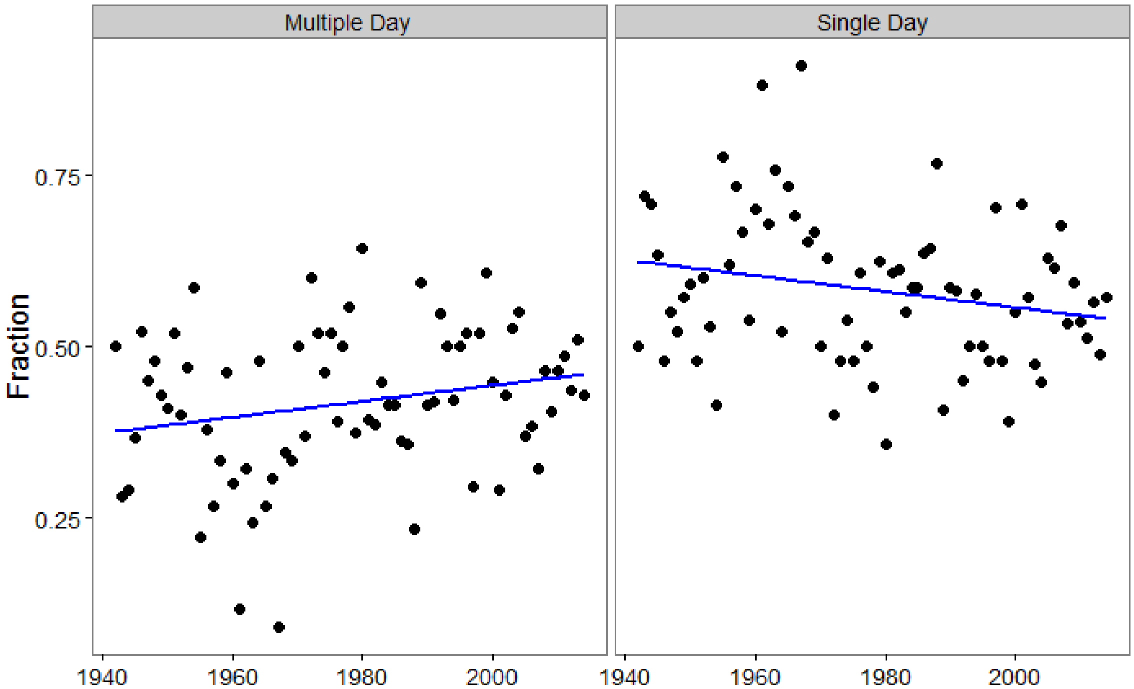

Figure 4.7: Annual fraction of multiple vs. single day rainfall events at Smith Creek, SK. The fraction of events falling as multiple day events has increased throughout the study period, whereas the fraction of single day events has decreased. Blue line represents significant linear trend ($p<0.05$)

Snow on ground data from Yorkton, SK was analyzed for maximum snow depth, duration of continuous snow cover, and the first snow free date from 1956 to 2012 (Figure 4.8; 2007 is missing). As expected from the decreases in snowfall and increases in temperatures (particularly in January and March), the maximum depth of snow has significantly decreased ($p<0.001$) at a rate of 8 cm/decade, which is 3 cm/decade faster than the decrease in snowfall. The rates of decline may differ due to the densification of the snowpack or increases in mid-winter melting due to the increasing temperatures. The duration of continuous snowcover has decreased ($p=0.058$), resulting in a significantly earlier first snow free date ($p=0.042$). On average, the duration of continuous snow cover is 128 days (~4.25 months), with the first snow free date gradually advancing from April 9th to March 26th. Subsequent snowfalls can occur after the first snow free date, but they tend to be small (<5 cm) and coverage short-lived.

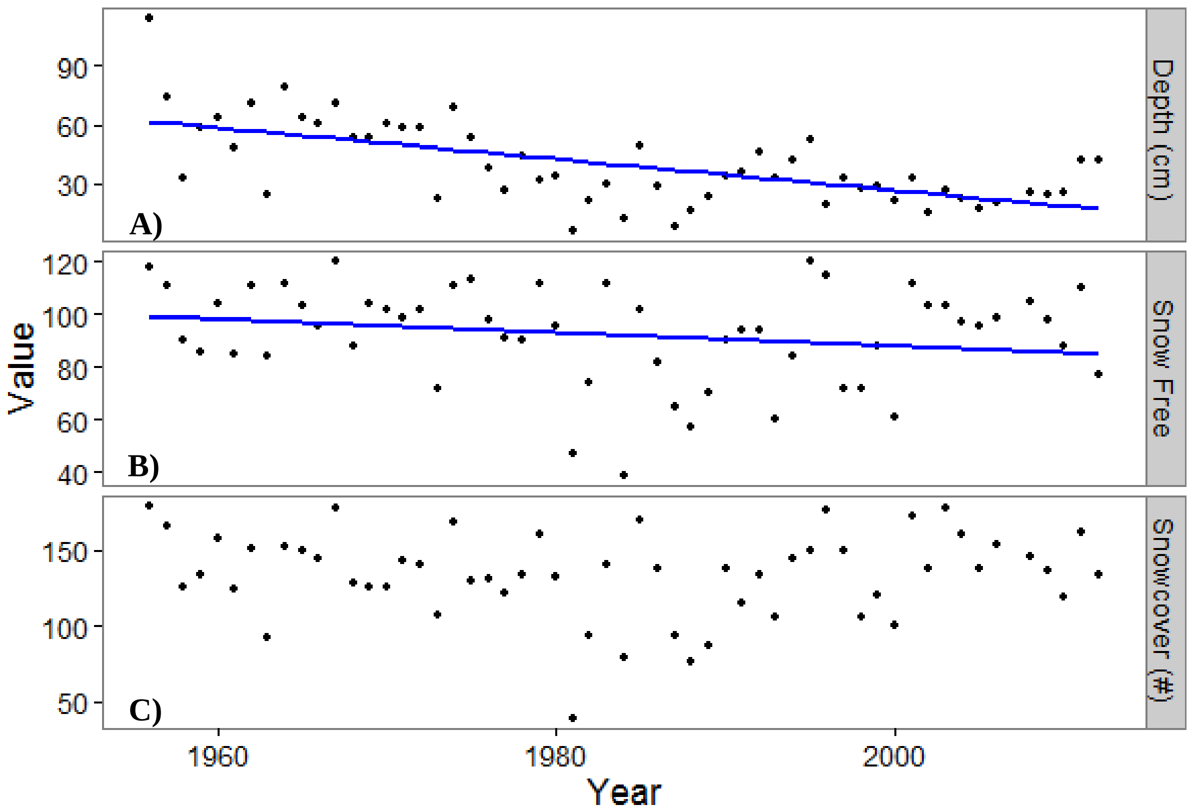

Figure 4.8: A) Maximum depth of snow per year (cm), B) First snow free date of the year represented by the Julian Date, and C) Duration of continuous snowcover days, 1956 to 2012 (missing 2007). Blue line represents significant linear trend at $p<0.05$.

4.4 Streamflow

Streamflows were analyzed for trends and changepoints for the period of 1975 to 2014 (note: 2014 is provisional data). Figure 4.9 shows the annual streamflow volumes and the runoff mechanism that produced streamflow (mixed refers to snowmelt and rainfall runoff mechanisms occurring simultaneously). Annual streamflow volume in the SCRB has increased by 14-fold (significant at $p < 0.001$) over the period of record. The average annual streamflow volume from 1975 to 2014 is 9715 dam^3, and changepoints were identified in 1994 and 2010 (Table 4.3) that increased the average annual volume. The annual average and peak streamflows were not found to be correlated to the PDO (R^2 = 0.329, R^2 = 0.291, respectively) or SOI (R^2 = 0.281, R^2 = 0.284, respectively), although it should be noted that the 40 years of available data mainly encompassed the positive phase of the PDO (1977 to 2007; St. Jacques et al., 2014). Streamflow volumes increased between 1995 and 2010, a period that includes the worst multi-year drought (1999-2004) on the Canadian Prairies (Bonsal and Wheaton, 2005). The second changepoint marked the beginning of one of the wettest years on record in the northern portion of the Canadian Prairies (Chun and Wheater, 2012), with the SCRB recording an annual rainfall volume of 498.8 mm in 2010 which ranked the 2nd highest annual rainfall amount since streamflow records started (1981 ranked 1st with 503.4 mm of rainfall). Annual rainfall in 2012 was 3rd highest with 477.6 mm. During the last changepoint, the Assiniboine Watershed experienced widespread flooding in 2011 and 2014, which included the SCRB. Furthermore, localized flooding in the SCRB was documented in 2012.

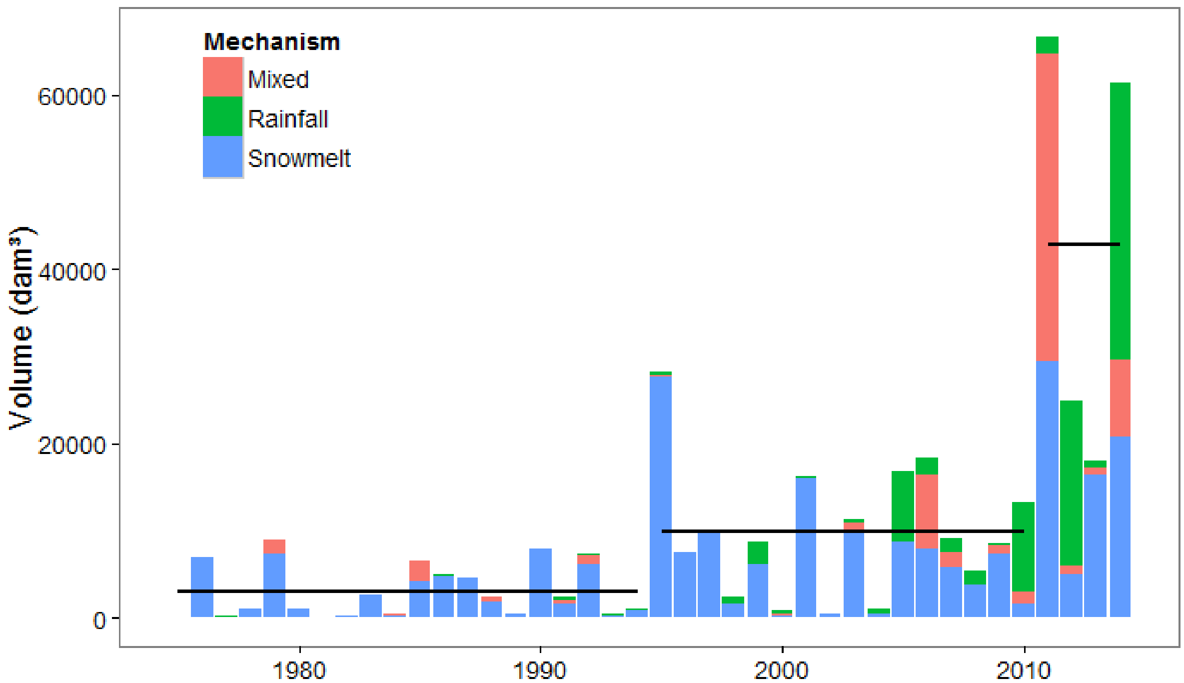

Figure 4.9: Annual streamflow volumes separated into varying runoff mechanisms for Smith Creek, SK from 1975 to 2014. Changepoints in the mean and variance (black lines) are identified in 1994 and 2010. NOTE: 2014 data is preliminary.

Not only has the volume of runoff increased within the SCRB, but the runoff mechanisms that produce streamflow have shifted throughout the study period as evidenced from variations in the increase in contributions from snowmelt, rainfall and mixed snowmelt-rainfall runoff since 1975 (Table 4.4 and Figure 4.9). Despite observed decreases in snowfall amounts, streamflow volume derived from snowmelt runoff increased 5-fold since 1975. However the increase in runoff mechanisms associated with rainfall was greater than that associated with snowmelt; mixed snowmelt-rainfall runoff volumes increased 34-fold and rainfall runoff volumes increased 150-fold over the same period.

Although streamflow volume derived from all forms of runoff increased and contributed to extensive flooding after the 2010 changepoint, the increases favoured rainfall-runoff mechanisms. For instance, snowmelt and mixed runoff volumes increased three-fold, and rainfall runoff volume increased 15-fold after the first changepoint and before the second changepoint. Prior to the first changepoint in 1994, snowmelt runoff accounted for 86% of the annual streamflow within SCRB, a value consistent with Gray and Landine's (1988) estimate for the Canadian Prairies of >80%. Since then, snowmelt runoff fractions have declined to 70% (1995 to 2010) and then 46% (2011 to 2014), while mixed and rainfall runoff have increased to 30% (1995 to 2010) and then 54% (2011 to 2014), suggesting a reversal of the basin from being snowmelt dominated to being rainfall-runoff dominated (Table 4.4, Figure 4.9). Changes in the nature of runoff continued after the second changepoint, with the additional increase in the fraction of streamflow derived from mixed snowmelt and rainfall runoff. The substantial changes in the contributions to streamflow from the runoff mechanisms after the second changepoint occurred in years with high early summer precipitation that produced rain-induced flooding in 2012 and 2014. Further, streamflow during the 2011 flood was dominated by mixed snowmelt-rainfall runoff as there were high amounts of rainfall that occurred near the end of the snowmelt period. This event marked a changepoint in the mixed snowmelt-rainfall runoff fraction of streamflow.

Table 4.3: Volume of streamflow at Smith Creek, SK derived from the different runoff mechanisms for 1975 to 2014. Note: 2014 data is preliminary.

	1975-1994	1995-2010	2011-2014
Annual Runoff	3,000	9,900	42,700
Snowmelt Runoff	2,600	7,210	17,880
Mixed Runoff	300	860	11,500
Rainfall Runoff	80	1,800	13,350

Table 4.4: Percentage of streamflow derived from the different runoff mechanisms at Smith Creek, SK for 1975 to 2014. Note: 2014 data is preliminary.

	1975-1994	1995-2010	2011-2014
Snowmelt Runoff	86%	71%	47%
Mixed Runoff	7%	6%	19%
Rainfall Runoff	7%	23%	34%

Since streamflow on the Canadian Prairies has been observed in the 20th century to be largely derived from snowmelt and typically peaks in April and ceases by May due to a lack of inputs (e.g. Gray, 1970), annual streamflow volume was separated into spring (March, April, May) and summer (June, July, August, September, October) time periods assess temporal changes. The volume of both spring and summer streamflow has increased significantly from 1975 to 2014 ($p < 0.001$; Figure 4.10), with two changepoints occurring for both spring and summer time periods (Table 4.5). Spring streamflow volumes experienced an 8-fold increase since the mid-1970's, whereas summer streamflow volumes increased 28-fold since the 1990's (omitting the 1975-1989 time period when streamflow was negligible), meaning that streamflow increased more substantially in the summer than the spring. On a monthly basis, streamflow volumes have increased significantly in April ($p = 0.01$), May ($p = 0.005$), June ($p < 0.001$), July ($p < 0.001$), August ($p < 0.001$), September ($p = 0.035$), and October ($p = 0.043$). March was the only month in which streamflow did not change ($p > 0.1$).

Table 4.5: Annual streamflow volume divided in spring (March, April, May) and summer (June, July, August, September, October) time periods based on changepoint analysis.

Spring (MAM)		**Summer (JJASO)**	
Changepoint Period	**Volume (dam^3)**	**Changepoint Period**	**Volume (dam^3)**
1975 - 1994	3085	1975 - 1989	37
1994 - 2010	8273	1989 - 2004	308
2010- 2014	26979	2004 - 2014	8548

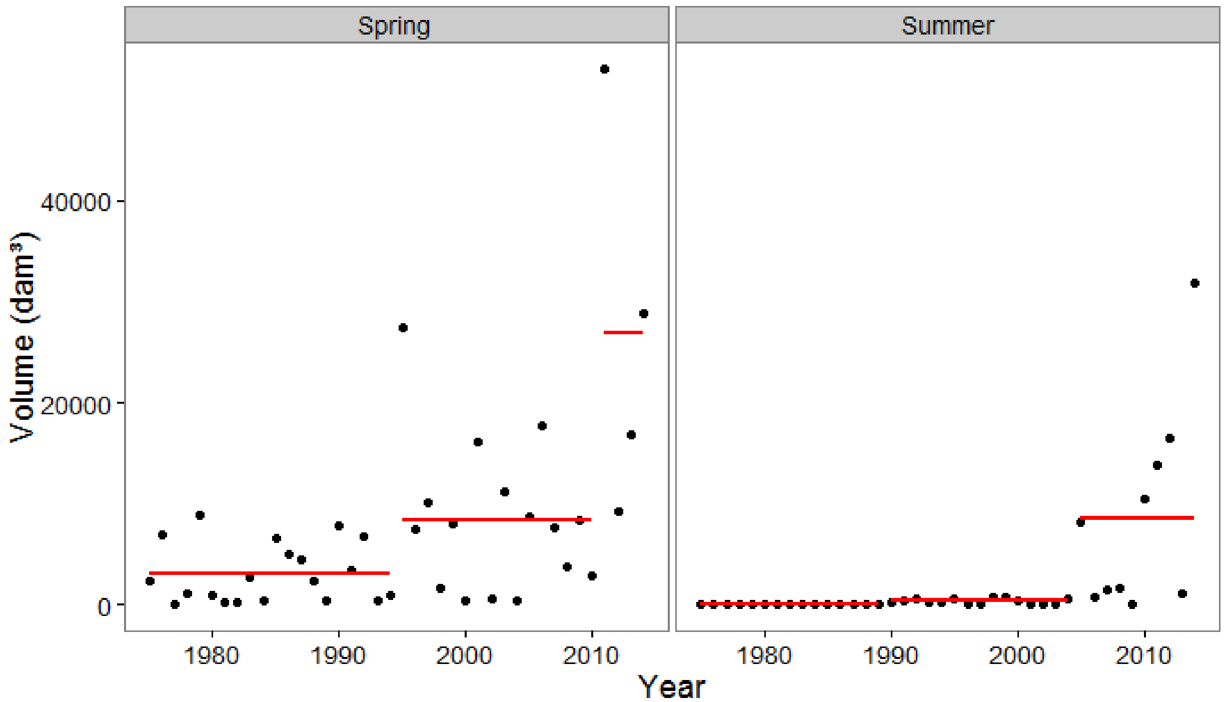

Figure 4.10: Annual streamflow volumes divided into spring (March, April, May) and summer (June, July, August, September, October) time periods. Red lines represent changepoints.

Runoff ratios were evaluated in order to gain greater insight into the role of changing precipitation in generating runoff within the SCRB. From 1975 to 2014, runoff ratios increased significantly ($p < 0.001$; Figure 4.11) and the period was marked by two changepoints which occurred in 1994 and 2010. The changepoints represent a shift in the mean annual runoff ratios from 0.019 prior to 1994, to 0.057 in 1995-2010, and then to 0.22 after 2011. These shifts were coincident with shifts in annual streamflow volume and amount to a 12-fold increase in runoff ratio over the period of record.

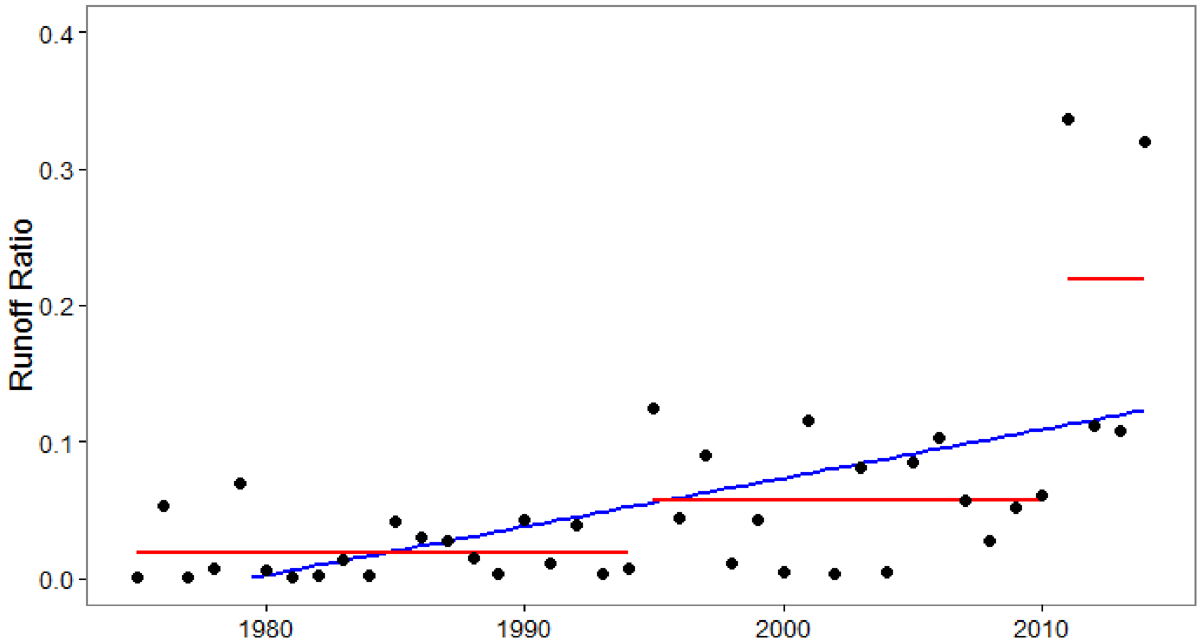

Figure 4.11: Runoff ratios calculated for Smith Creek, SK for 1975 to 2014. Blue lines represent significant linear trend ($p<0.05$) and red lines signify changepoint time periods.

Figure 4.12 depicts the maximum, minimum and mean hydrographs of the three changepoint time periods of annual streamflow. From 1975 to 1994, streamflow in SCRB was dominated by snowmelt runoff and typically receded back to zero flow after the completion of snowmelt in spring. Negligible runoff occurred during the summer months. From 1995 to 2010, streamflow duration expanded into the summer months due to rainfall-runoff mechanisms and rain-on-snow contributed to generating higher runoff peaks in the spring period. The largest peak flow during this period (1995) was caused by widespread spring flooding resulting from high fall soil moisture levels coupled with high amount of snowfall (Government of Manitoba, 2013). Streamflow was first sustained throughout much of the summer in 2010. Hydrograph shapes for the period of 2011 to 2014 are substantially different from other years. During this period streamflow still peaked in spring due to snowmelt and mixed runoff, but there were second peaks that occurred in the summer months and were not driven by snowmelt. Double peaks occurred in 2012 and 2014, with the rainfall driven peak being larger than the snowmelt peak in both years.

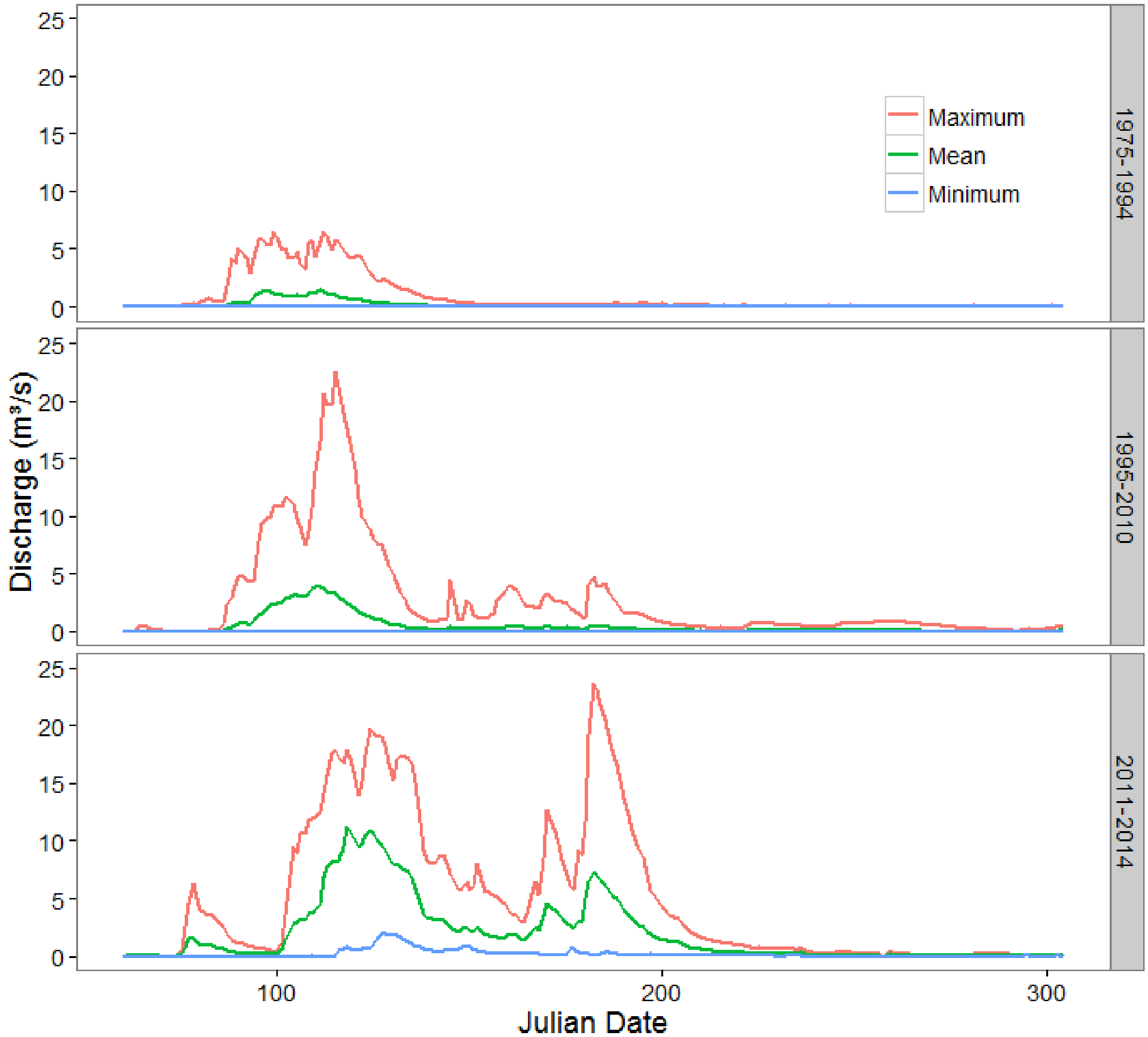

Figure 4.12: Maximum, minimum and mean daily discharges for the three changepoint time periods identified in streamflow volume records for Smith Creek, SK. Note: 2014 data are preliminary.

Not only has the volume of runoff and streamflow increased within the SCRB, but the annual peak discharge has also increased substantially (Figure 4.13). From 1975 to 2014, annual peak discharge at Smith Creek increased significantly ($p = 0.006$), with a changepoint occurring in 1994. Prior to the changepoint, mean annual peak discharge was 3 m^3/s, which tripled to 9.2 m^3/s afterwards. There is greater uncertainty in highest peak discharge values because the culvert at the gauging station was inundated during peak flow in 1995, 2011 and 2014. Although the diameter of the culvert at the gauging station has not been altered, the roadway going over the culvert has been built up, increasing the culvert length. The higher road height might allow for the development of a deeper headpool upstream of the culvert, resulting in increased flow rates through the culvert during very high flows (Smith, 1985; see section 3.2.5 - Streamflow). Impacts of these alterations are only likely during extremely high flows wherein the culvert is overtopped and upstream ponding occurs. In these cases, a whirlpool forms over the inlet – whirlpools were observed in 2011 and 2014 (see Figure B.4 and B.7). Despite these potential restrictions on high streamflows, the alterations to the culvert length and road height cannot fully explain the tripling of the annual peak discharge after 1994 or 12-fold increase in runoff ratios.

The timing of the annual peak discharge was found to be significantly increasing (p = 0.013; Figure 4.14), meaning the annual peak flow is occurring later in the year. A changepoint occurred in 2009, after which the annual peak flow date changed from April 8th to May 19th. The changepoint corresponds closely to the second changepoint in annual streamflow volumes (2010) where annual peak flow dates in 2010, 2012, and 2014 were rainfall driven and occurred on or after June 19th.

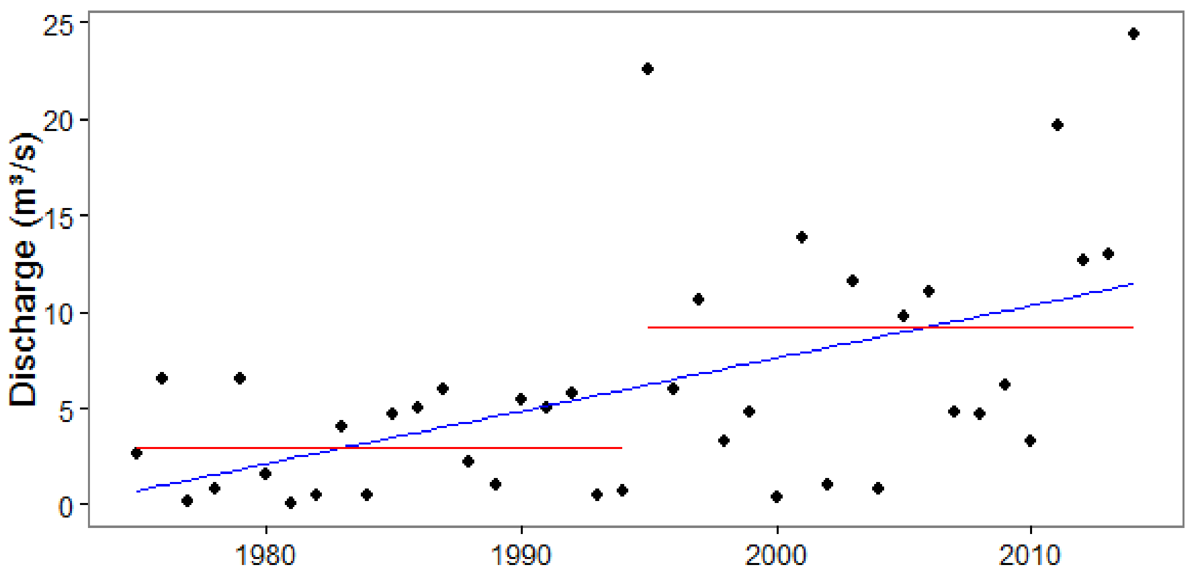

Figure 4.13: Annual peak discharge from the outlet at Smith Creek, SK from 1975 to 2014.Note: 2014 data is preliminary. Red lines show changepoints, blue line shows significant linear trend (p<0.05).

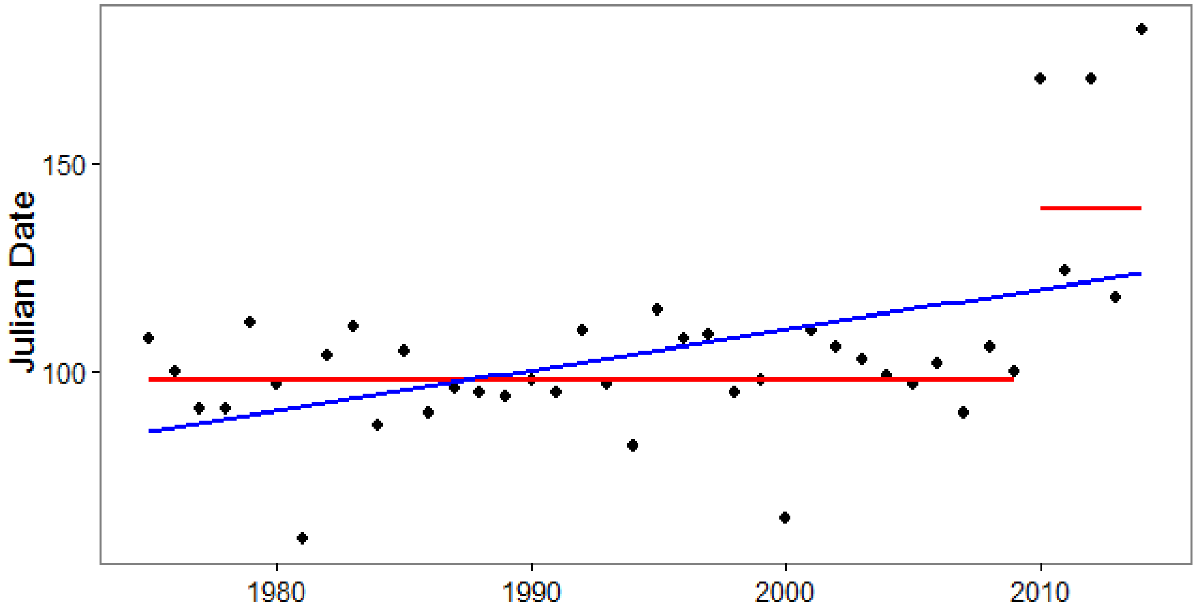

Figure 4.14: Date of annual peak discharge at the outlet of Smith Creek, SK from 1975 to 2014. Red line shows changepoints, blue line shows significant linear trend ($p<0.05$).

Annual peak flow has not always been derived purely from snowmelt processes (e.g. 2010, 2011, 2012, 2014). In order to eliminate the recent uncharacteristic rainfall driven peak flows in the SCRB, spring peak streamflows were examined to focus on the changes in timing and magnitude of snowmelt driven peak flow events. The timing of spring peak flow has not changed ($p > 0.1$) between 1975 and 2014, with an average date of April 9th. The magnitude of the spring peak flow has increased significantly ($p = 0.004$) with a changepoint occurring in 1995 that over doubles spring peak flow from 3 m^3/s to 8.3 m^3/s.

The duration of flow can be seen in Figure 4.15 and is defined as the number of days where flow is greater than 0 m^3/s. The duration of flow has gradually and significantly increased during the period of record ($p < 0.001$), with a changepoint occurring in 1990. The mean number of days per year with flow is 117. The average duration of flow was 73.5 days prior to the changepoint and 146.7 days after it.

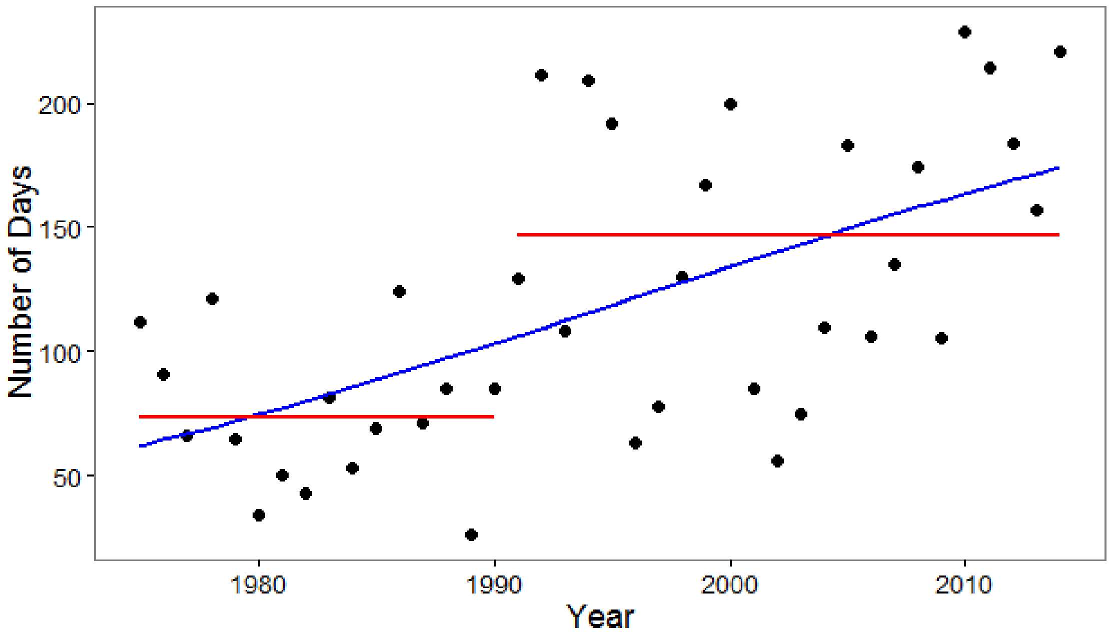

Figure 4.15: Annual flow duration at Smith Creek, SK from 1975 to 2014. Red line shows changepoints, blue line shows significant linear trend ($p<0.05$). Note: 2014 data is preliminary.

4.5 Observed Flooding

Since streamflow records started in 1975 for the SCRB, four major flood events have occurred in the Assiniboine Watershed: 1976, 1995, 2011 (Government of Manitoba, 2013), and 2014. Streamflow records from the SCRB identified three flood years (1995, 2011, 2014) with no indication of flooding occurring in 1976 (see Figure 4.9). In addition to the three widespread flood events, localized flooding occurred in SCRB in 2012 in which the RM of Churchbridge declared a state of emergency (Ahmad, 2012). All four events were driven by varying climatic conditions and were derived from varying runoff processes. A summary of each flood year can be found in Table 4.6.

Table 4.6: Summary of the four flood years recorded in SCRB from 1975 to 2014 (2012 was a localized flood). Table includes annual data on streamflow volume, peak discharge and the percentage of streamflow derived from each runoff mechanism.

	1995	2011	2012	2014
Volume (dam^3)	28,140	66,743	24,965	61,353
Peak Discharge (m^3/s)	22.6	19.7	12.6	24.4
Peak Date	Apr 25	May 4	June 19	July 1
Snowmelt	99%	44%	20%	33%
Mixed Rainfall-Snowmelt	0%	53%	4%	15%
Rainfall	1%	3%	76%	52%

The flood of 1995 was the third largest streamflow event in the SCRB. High rainfall amounts in late summer and early fall in 1994 produced wet conditions (Government of Manitoba, 2013) going into winter with high snowfall amounts. Rainfall depths in August to November 1994 totalled 179 mm, with 130 mm of that falling in August. Localized runoff and direct precipitation from these summer and fall rains may have acted to increase water storage in depressions. This large volume of rainfall may have increased the soil moisture content of soils in the basin at the time of fall freeze-up, which has a large influence on reducing infiltration, and therefore increasing runoff, during the subsequent spring melt. For instance, infiltration to frozen soils in 1995 may have been restricted by the formation of an ice lens over frozen ground as in November 1994 when 5 mm of rain fell onto a shallow snowpack. The snowfall over the winter of 1994-1995 was 180 mm snow water equivalent (SWE), which was the highest snowfall amount since streamflow records started in 1975. Temperatures warmed in early March 1995 and initiated snowmelt. On March 16, 1995, mixed precipitation consisting of 7 mm of rain and 10 mm SWE of snow fell. Snow was recorded on the ground until April 20th, a week after another mixed precipitation event occurred, consisting of 5 mm of rain and 12 cm of snow. The total streamflow volume in 1995 was 28,140 dam^3 and was the highest on record at that time. Peak flow was estimated at 22.6 m^3/s and occurred on April 25th.

Widespread flooding in the spring of 2011 was caused by more extreme conditions than in 1995. Annual rainfall in 2010 was 499 mm (ranked 2nd highest since streamflow records started in 1975), 175 mm over the long term average and much of this occurred late in 2010, wetting up the basin and soils going into fall and winter. 2010 was the first year on record where peak flow in Smith Creek was caused by rainfall and was also the first year where streamflow was sustained throughout the summer. On October 23-25 2010, two days before the first snowfall, 43 mm of rain fell, saturating the surface layer of soils just before freeze-up. Snowfall over the winter from 2010 to 2011 was close to average, at 130 mm SWE. Snowmelt was initiated by a rain-on-snow event on March 16, 2011 with 8 mm of rain falling; snowmelt was not completed for a further ~6 weeks. Streamflow started April 8-10th 2011, and culverts began overtopping on April 16th. Many culvert gates were closed in the basin at this time, and many culverts that were not equipped with gates were boarded up. A flow direction reversal was observed on April 19th near the centre of SCRB. Water normally flowed through a culvert into the main-stem of Smith Creek but as Smith Creek rose, water was forced back over the road in the opposite direction to the original flow. The SCRB was partially snow-covered until May 2nd. Peak flow occurred on May 4th and was estimated by WSC to be 19.7 m^3/s but this flow was restricted by the rate of flow from a grid road-dammed pond through the submerged culvert immediately downstream of the gauge. Two major rainfall events occurred near the end of the snowmelt: 29 mm on April 28th and 35 mm on May 9th. The annual discharge volume in 2011 was 66,746 dam^3 which vastly exceeded the 1995 flood.

The localized flood of 2012 differed substantially from the floods of 1995 and 2011, which consisted of snowmelt or mixed snowmelt and rainfall runoff floods occurring in April or May. The 2012 flood occurred in mid-June after the cessation of snowmelt derived streamflow and was the first major flood produced by rainfall-runoff recorded in the basin, following the first high flow due to rainfall-runoff in 2010. Snowfall over the 2011-2012 winter was well below average at 87 mm SWE, and snowmelt occurred early and modestly with snowmelt derived streamflow peaking at 6.4 m^3/s on March 19th. Streamflow within the Smith Creek basin did not cease until Aug 31st due to 478 mm of rain in the summer, which was 151 mm over the long term average. Rainfalls were concentrated in April, May, and June with 50.6 mm, 114.5 mm, and 158.9 mm, respectively. The previous year's heavy snowmelt and the 2012 snowmelt and spring rainfall filled much of the depressional and soil storage in SCRB early in the summer. The flood in 2012 was triggered by an intense and spatially variable storm that was reported by locals to have delivered 100 to 150 mm of rainfall in parts of the SCRB. The SC MET station recorded 74 mm from June 5-15th followed by 52.5 mm of rainfall within this 24-hr storm period on June 17-

18th. The intense 24 hour storm contributed high amounts of runoff to depressions and then fill and spill or direct drainage to the stream, causing the first ever rainfall-driven flood event on record in Smith Creek. The annual discharge volume in 2012 was 24,965 dam^3 which is just shy of the 1995 flood. Peak flow occurred on June 19th and reached 12.6 m^3/s.

The widespread flood in 2014 was similar to 2012, yet it was unique given that it was not only a rainfall driven flood event, but was the largest flood since records started in the SCRB. The flood of 2014 occurred at the end of June and early July from heavy multi-day rainfall amounts that occurred well after snowmelt derived streamflow ceased. Snowfall over the 2013 – 2014 winter was around average, with 107 mm SWE recorded. Streamflow derived from snowmelt peaked at 17.8 m^3/s on April 28th and never ceased, but declined to ~ 0.2 m^3/s by mid-June. From June 27th to June 30th a large frontal storm that blanketed southeastern Saskatchewan and moved into Manitoba dumped varying amounts of rainfall in the region. Rainfall depth measurements from in and around the SCRB ranged from 110 mm to 200 mm over the 4 day period (in correspondence with Neil Mehrer). The intense multiple day storm contributed high amounts of runoff to depressions and streams within the basin. The gauging station at the outlet of Smith Creek failed on June 30th due to the formation of a grid road-dammed pond behind the outlet that drowned out the gauging station. Technicians from the WSC took a manual flow reading on July 1st which read 24.4 m^3/s ~100 m downstream of the culvert and gauging station. This measurement is assumed to be close to the peak based on reports from local residents and inspection of available air photographs. The annual discharge measurements are preliminary as the WSC has not yet conducted quality control on the data. It is important to note that the water did not overtop the roadway. The magnitude of the total annual discharge was initially found to be 61,353 dam^3. The 2014 was not only the largest flood on record in regards to peak flow, but it was a flood driven solely by rainfall which was once thought to account for less than 20% of the annual runoff on the Canadian Prairies.

Chapter 5 : Discussion

The climate, land use and hydrological regime of SCRB have changed dramatically in the last half-century. Changes to streamflow and drainage were large and much greater than changes to climate. Critically, the annual volume of precipitation to the basin has not changed significantly over time, but important changes were observed in how precipitation is delivered, which impact streamflow.

Temperature increases in the SCRB were largest in winter and early spring which agrees with other research on the Canadian Prairies (Gan, 1998; Zhang et al., 2000; Bonsal et al., 2001; Millet et al., 2009). Maximum daily temperatures averaged over the year increased 1.2 °C whereas minimum daily temperatures did not change significantly, which differs from other regionally based research showing larger increases in annual minimum than maximum temperatures (Zhang et al., 2000; Millet et al., 2009).

The annual precipitation in the SCRB has not changed significantly in contrast to other published records on the Canadian Prairies which show increasing trends (Akinremi et al., 1999; Millet et al., 2009). Despite a lack of changes to the annual precipitation, there have been significant increases in rainfall and decreases to snowfall. A 3.5 °C increase in March temperature since the early 1940's has instigated a quadrupling of the March rainfall fraction. Shook and Pomeroy (2012) also identified a shift in the spring (and fall) precipitation towards rainfall than snowfall. The increase in rainfall in the SCRB was largely concentrated in May and June, accounting for 28 and 34 mm, respectively. In the summer months the number of multiple day rainfall events has increased by 50%, most of which are 2-days in duration, with no significant changes in the magnitude of events.

The decrease in annual snowfall amounts coupled with increased winter and early spring temperatures have brought on a decrease in the maximum snow depth as well as an advance of the first snow free date by two weeks to March 26th. Maximum snow depth has declined at a faster rate than snowfall amounts and is most likely attributed to increased winter temperatures which can increase the densification of snow as well as the occurrence of mid-winter melts (Burn et al., 2010). Despite a two week earlier snow free date, the timing of the spring peak flow has not changed significantly (April 9th). This result slightly differs from other research on the

Canadian Prairies that has identified snowmelt occurring earlier by examining spring peaks in the streamflow records (Burn, 1994; Gan, 1998; Burn et al., 2010).

Despite decreases in snowfall depth, the volume of streamflow derived from snowmelt runoff has quintupled. Increasing March rainfall may have increased the frequency of occurrence or spatial extent of ice layers at the base of the snowpack due to rain-on-snow events. Ice layers restrict the infiltration of snowmelt water into frozen soils (Gray et al., 2001) and can cause almost all snowmelt to form runoff. Additionally, the increase in snowmelt runoff in spring may in part be due to increased rainfall in October. If soils freeze when wet or saturated, the snowmelt generated in the following spring can be limited or restricted from infiltrating (Gray et al., 1986).

The increases in rainfall amounts have largely been concentrated in May and June, and combined with a 50% increase in the frequency of multiple day rainfall events in the summer, it has almost certainly contributed to the increase in the amount of mixed and rainfall runoff observed at the basin scale. As seen in Figures 4.10 and 4.12, the existence of streamflow has consistently extended into the summer months, particularly after 1990 when the volume of streamflow in the summer months (Table 4.5) and the duration of streamflow (Figure 4.14) increased substantially. Streamflow volumes after 2004 in the summer months experienced a 28-fold increase (Figure 4.10), illustrating the increasing incidence of rainfall runoff for streamflow generation in the SCRB. The concentration of rainfall shortly after the snowmelt period when the basin is still relatively wet likely played a role in the observed increases in rainfall runoff. From 2011 to 2014 (Figure 4.12), a second peak emerged in the hydrographs during the summer months. Other research on the Canadian Prairies has similarly identified that streamflow will primarily shift further into the rainy season due to an increase in the rainfall fraction of precipitation (Burn et al., 2010). This has been observed in the SCRB, particularly in 2010, 2012 and 2014 when peak flows during those years were derived from rainfall runoff events.

Abrupt changes in annual streamflow volume and runoff ratios in 1994 and 2010 do not directly correspond with any shifts in precipitation characteristics in the SCRB. The observed changes to the character of precipitation were too gradual and small to cause the 14-fold increase in streamflow volumes, 12-fold increase in runoff ratio, and tripling of peak discharge over the period of record. Streamflow records end shortly after the 2010 changepoint, with a need for re-evaluation as a longer record becomes available.

Though flow frequency analysis was not conducted, it is clear that the hydrological regime of Smith Creek is non-stationary. Research in the prairie region has shown that not only do changes in precipitation influence streamflow, but so do anthropogenic changes (Miller and Nudds, 1996; St. Jacques et al., 2010; Sheikh and Bahremand, 2011) such as water diversion, storage and consumption as well as land modifications. In the SCRB, over 58% of the ponded area has been lost due to drainage, reducing the wetland storage volumes by 79%. The drainage channels that have been created in SCRB are generally well-engineered, allowing for little to no residual depressional storage (Mr. Don Werle – personal communication). Farmers in the basin have openly stated that they amplify their drainage efforts in flood years to reduce the amount of water sitting on the landscape and associated economic losses (Brunet and Westbrook, 2012).

Modelling studies have concluded that wetland drainage increases annual and peak daily flows, as well as the magnitude and frequency of flooding (Gleason et al., 2007; Yang et al., 2010; Pomeroy et al., 2014). Streamflow generation from Canadian Prairie runoff is controlled by depressional storage (Shook et al., 2015) and draining depressions increases the contributing area to streamflow (Brannen, in review). Comparable results were presented in Miller and Nudds (1996) who concluded that landscape alteration caused an increase in runoff in unregulated rivers, as there were no changes to the amount of precipitation. Ehsanzadeh et al. (2014) also examined the SCRB streamflow records from 1975 to 2005 and found significant upward trends in cumulative and peak annual flow and were associated with drainage. A near-significant doubling of the runoff ratios in 1991 was also noted (Ehsanzadeh et al., 2014) which differs from this study due to the differing lengths in streamflow records and source of precipitation data. It was further discussed in Ehsanzadeh et al. (2014) that the impact of drainage may intensify streamflow generation in smaller scale watersheds, whereas the impact may be muted in larger ones.

The increase in drainage channel length between 1958 and 2000 was four fold, whereas an eight-fold increase was observed from 1958 to 2009, suggesting an increase in the rate of drainage during the 21st century. Ponded areas in SCRB were reduced by 14%, drainage channel lengths increased by 185% and wetland storage volumes declined 20% from 2000 to 2008/2009.

The 12-fold increase in runoff ratios from the 1970s to the current decade likely reflects factors such as changing precipitation delivery, infiltration capacity, contributing area, depressional storage, and connectivity of the basin. Whilst spring infiltration rates may have declined due to increases in March and October rainfall and the greater frequency of multiple day rainfall events in May and June can increase summer runoff, the marked increase in late spring and summer flows due primarily to increasing mixed and rainfall runoff is disproportionate to climate change impacts. Over half the depressions were drained in the last 56 years leaving only 1/5th of the maximum wetland storage volume and increasing drainage channel lengths eight-fold, resulting in a decline in depressional storage capacity and increase in contributing area through increased connectivity. It is therefore highly likely that the increase in runoff ratios, along with annual streamflow volumes and peak discharge, is partly driven by enhanced drainage. Evidence of drainage affecting streamflow in the SCRB was first known to be documented on November 6, 1980 by WSC technicians who noted in site visit notes that streamflow was a result of ditching or slough drainage.

Changes in agricultural land use in the SCRB include substantial increases in crop land and extensive adoption of zero and conservation tillage practices. On the Canadian Prairies, the conversion of land use to crop land decreases infiltration in the spring (van der Kamp et al., 2003; Pomeroy et al., 2010) due to the destruction of macropores (van der Kamp et al., 2003). At the same time, shorter plant heights result in decreased snow accumulation due to increased blow snow sublimation (Pomeroy et al., 2010). Results from a modelling study in the SCRB found that spring streamflows were relatively unaffected (-2%) under complete conversion to crop land (Pomeroy et al., 2010). Runoff differences under conservation and zero tillage practices have been shown to have little effect during the snowmelt period, yet can effectively reduce runoff during the growing season (Elliott et al., 2001; Tiessen et al., 2010). Personal communication with a long-term farmer in the SCRB (Mr. Don Werle) regarding the impacts of tillage practices coincides with the published research. Zero till practices are beneficial during dry periods as the stubble remaining on the field acts to trap snow, increasing infiltration and soil moisture in the spring and decreases moisture losses throughout the summer. But during a deluge, substantial moisture can be retained in the soils under zero till practices. Thus, it is likely changes in agricultural land use have had little effect on the observed increased streamflow in the SCRB, and instead may have acted as a buffer through reducing runoff during the summer months when the largest changes in streamflow are observed. Further diagnostic hydrological process studies are needed to gain a better understanding of how land use change, including drainage, influence streamflow in this and similar basins.

The results from this study confirm and help explain many findings of recent research conducted on the Canadian Prairies. Ehsanzadeh et al. (2012a, 2014) suggested that changes in precipitation inputs would result in sudden changes to streamflow, which appears to be supported here in that a dramatic increase in streamflow volume occurred with changes in the delivery but not the annual volume of precipitation. The disproportionate response of streamflow to such changes may in part by due to the dynamic nature of the contributing area (Ehsanzadeh et al., 2012a). Many studies have shown the importance of depressional storage effects on streamflow generation (Gleason et al., 2007; Yang et al., 2010; Shook et al., 2015) which affects the shape and slope of the flow frequency curve (Ehsanzadeh et al., 2012b) and such non-stationary effects may be becoming apparent in SCRB as depressional storage declines with drainage. Ehsanzadeh et al. (2014) also found increased streamflow and runoff ratios for SCRB in their coarse-scale, regional analysis. Yet the lack of long term trends in the Assiniboine River and across the Canadian Prairies in Ehsanzadeh et al. (2012a, 2014) is not supported by the data and detailed analysis for SCRB shown here, where there are strong and statistically significant trends for increasing discharge. The differences among these studies may be the inclusion of regulated rivers in Ehsanzadeh et al. (2012a, 2014), which may mask the ability to detect increased runoff. Unfortunately most unregulated creeks and rivers in the region are ungauged, and so it is not possible to assess at broader, regional scale whether unregulated rivers show non-stationarity in flows due to changes in basin hydrology in the same manner as this study has shown for SCRB.

Chapter 6 : Conclusion

The hydrological regime of the SCRB has changed substantially between 1975 and 2014. Drainage has increased in the basin over the last 56 years, reducing the ponded area extent by 58%, increasing drainage channel length 780% and reducing the maximum wetland storage volumes by 79% between 1958 and 2008/2009. The climate change observed within the basin is one of warming, with air temperatures increasing substantially in late winter and early spring. In response, the rainfall fraction of total precipitation has gradually increased, with the greatest shift from snowfall to rainfall events in March. Snowfall amounts have decreased, resulting in a considerable decline to maximum snow depth and an earlier snowmelt by two weeks. The frequency of multiple day rainfall events has increased by 50% and this may reflect greater frontal precipitation rather than convective precipitation mechanisms in the summer. Increased rainfall has predominately occurred in May and June, which is shortly after the snowmelt season ends and the basin is still relatively wet.

Streamflow volumes, runoff ratios, and peak discharge have increased dramatically since 1975, with large shifts occurring in 1994 and 2010. Despite decreases in snowfall amounts, the volume of streamflow derived from snowmelt has increased. However, vastly increased discharge contributions from rainfall and mixed rainfall-snowmelt runoff processes have changed the fraction of streamflow derived from various runoff mechanisms; snowmelt runoff fractions decreased and rainfall runoff increased. Mean annual peak discharge and runoff ratios tripled after 1994, and runoff ratios then almost quintupled after 2010. Flow duration experience a doubling after 1990, with substantial increases in summer streamflow occurring at the same time. Summer streamflow further increased 28-fold after 2004, with annual peak flows occurring from rainfall runoff in 2010, 2012, and 2014.

This study cannot attribute the dramatic shift in SCRB hydrology to any single cause. The substantial but gradual changes to the character of precipitation cannot fully explain the 14-fold increase in streamflow volumes, 12-fold increase in runoff ratios, transformation from snowmelt dominated to rainfall dominated runoff and tripling of peak discharge. This rapid shift in hydrology is likely due to a non-linear or threshold-like response to combinations of a changing climate, exacerbated by changes in agricultural land use and recent increases in drainage. Therefore, it is critical to consider the influence that drainage exerts on prairie basin hydrology under a changing climate. Further diagnostic investigation using process hydrology simulations is needed to fully explain the mechanisms behind the observed regime changes.

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Appendix A – Precipitation Adjustments

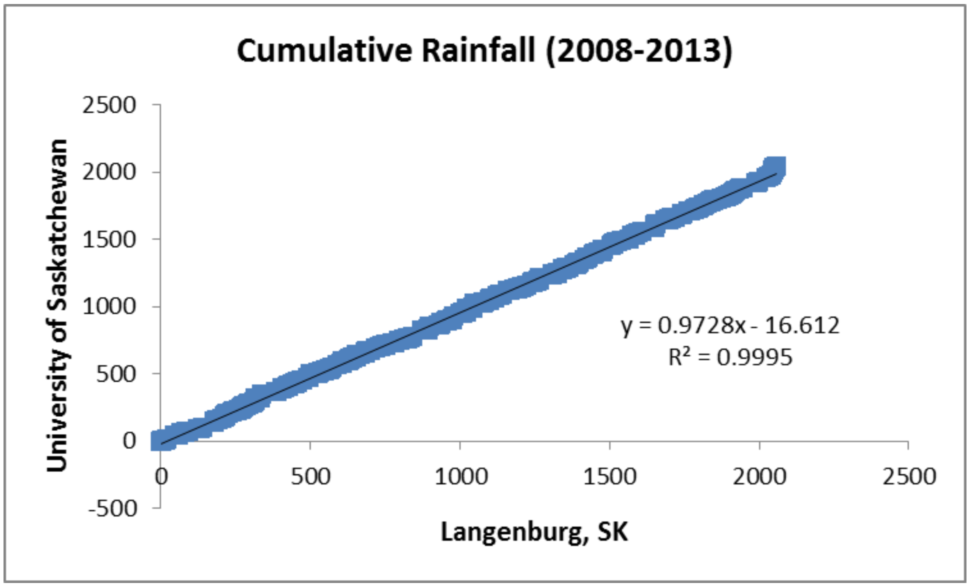

Figure A.1: Cumulative rainfall between Langenburg and University of Saskatchewan meteorological sites from 2008 to 2013. Double mass curve with linear regression.

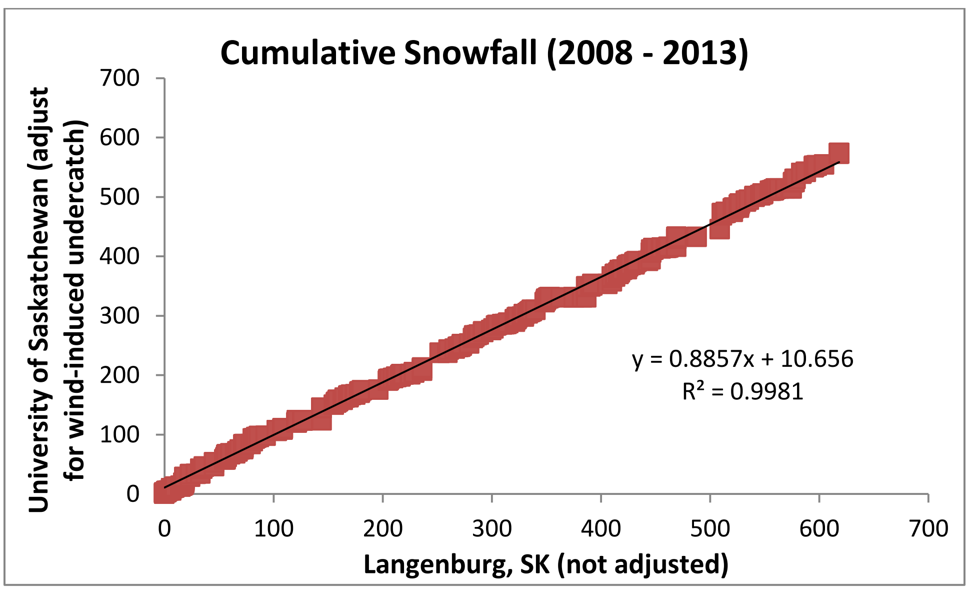

Figure A.2: Cumulative snowfall between Langenburg and University of Saskatchewan meteorological sites from 2008 to 2013. Double mass curve, with linear regression.

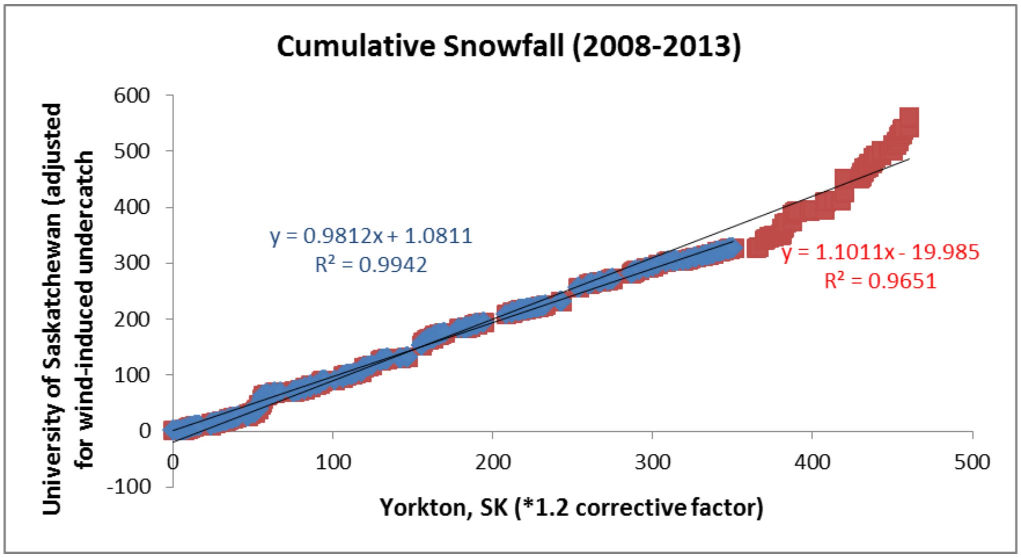

Figure A.3: Cumulative snowfall between Yorkton and University of Saskatchewan meteorological sites from 2008 to 2013. The red values and regression equation are for the entire data set (2008 to 2013), whereas the blue is the portion of the dataset used to derive a regression equation for this study (omits change in slope).

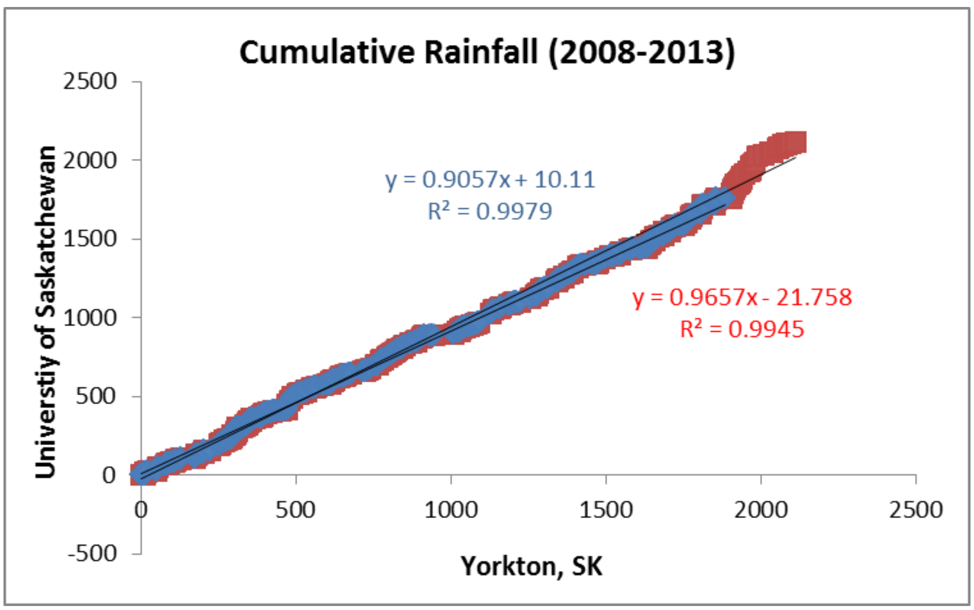

Figure A.4: Cumulative rainfall between Yorkton and University of Saskatchewan meteorological sites from 2008 to 2013. The red values and regression equation are for the entire data set (2008 to 2013), whereas the blue is the portion of the dataset used to derive a regression equation for this study (omits change in slope).

Appendix B – Study Site Photographs - SCRB

Figure B.1: Aerial photographs from Smith Creek Research Basin on April 28th, 2011. [Right] An area of SCRB where no drainage has occurred and the surface runoff generated is stored in the depressions on the landscape. [Left] An area of SCRB where wetlands have been drained and the surface runoff is no longer stored on the landscape, but follows the drainage channels to ditches or streams. Photo Credit: Ducks Unlimited Canada.

Figure B.2: Meteorological station set up in the SCRB in July 2007 and was dismantled in October 2013 (Photo Credit: Logan Fang).

Figure B.3: Hydrometric station at the outlet of SCRB run by Water Survey of Canada, April 22, 1995. Photograph taken 3 days prior to peak flow during the flood of 1995. (Source: Banga, 1996)

Figure B.4: Photograph looking upstream of the outlet of SCRB. Taken May 4, 2011 during a rain-on-snow flood event (1 day before peak). Photo credit: Nicole Seitz.

Figure B.5: Photograph taken June 21, 2012, 2 days after peak flow during the first ever rainfall driven flood event in the SCRB. Photo credit: Chris Marsh.

Figure B.6: Air photograph showing WSC gauging station at Smith Creek, SK on July 2, 2014, approximately 1 day after peak flow. Photo credit: Ducks Unlimited Canada.

Figure B.7: Outlet of SCRB looking upstream on July 4, 2014 approximately 3 days after record breaking peak flow occurred from rainfall. Photo credit: Stacey Dumanski.

Figure B.8: Photograph taken on April 13, 2008 (2 days prior to peak flow). Shown as a reference photograph for the outlet of the SCRB under average flow conditions during spring freshet. The culvert and height of the hydrometric station can be seen in this photo. Photo credit: Logan Fang.

Appendix C – History of WSC Gauging Station

The Smith Creek gauging station (near Marchwell, station number: 05ME007) was established in 1974 by the Water Survey of Canada (SWC). It is located ~3.2 km north of Marchwell, SK where a culvert has been located going through the road (2.44 m in diameter). A 1.5 m diameter by 3.0 m tall stave well with a metal walk-in shelter was installed 10 m south of the culvert on the upstream side of the road (Figure C.2). A metal plan walkway from the road embankment also exists. An intake pipe leads from the well to the creek bed to allow the well to mimic the stage of the creek.

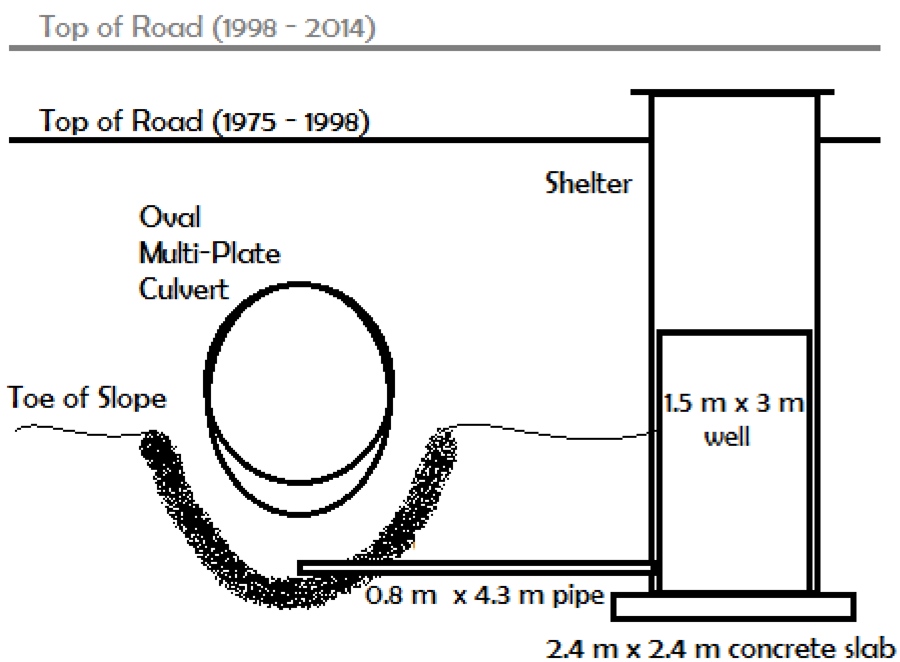

Figure C.1: Sketch of the Smith Creek gauging station looking downstream (adapted from public records at the Water Survey of Canada)

Continuous stage levels were recorded in the stilling well. Initial instrumentation to record stage included a float activated Steven's A Type Chart Recorder and recorded in imperial units until 1977 when it was changed to metric. A staff gauge was used as a reference. In 1987, an electric tape gauge (ETG) was installed and was used as a reference. In 2001, a float activated Amasser PDAS II logger was installed and referenced to the ETG. The float activated Steven's Type A recorder was also kept running as a backup until 2006 due to equipment and programming issues with the Amasser logger for the first 3 years of being installed. In

September of 2007, a new HDR Amasser logger equipped with GOES telemetry and separate Amasser shaft encoder were installed and used until the end of 2014. Due to the flood in 2014 which submerged the stilling well and shelter (including the equipment), the gauging equipment is expected to change from a stilling well to a pressure sensor system (via personal communication with Jeff Woodward).

Missing stage data existed throughout the records. Infilling techniques included using temperature, precipitation, and data from nearby gauging stations to best fill in the missing gaps and determine if and when melting and streamflow would most likely occur. A majority of the meteorological data was obtained from Yorkton, SK, with some use of the Langenburg, SK stations, both run by Environment Canada. Use of other gauging stations included Stony Creek and Jumping Creek, both monitored by the WSC. Corrections were also applied to the water level data and were based on reference checks on water levels 2-4 times a year. Corrections were identified by comparing reference gauges, such as the staff gauge and ETG, to the chart or digital recorders. Over the streamflow season, corrections to the water levels are linearly interpolated between reference checks and were typically off by millimetres, but could reach up to a couple centimeters.

Although stage was continuously recorded during the season, manual discharge measurements were taken ~12 times per year, with almost half of those measurements reading zero flow, on average. Discharge measurements were taken at various locations, with a majority being within the culvert or ~200 m downstream from the culvert at a rock wading section. If conditions didn’t allow, discharge measurements were taken at other locations, including upstream of the gauging station. Details on all of the locations were not immediately available. In personal communication with Jeff Woodward from the WSC, manual discharge measurements were taken using a price-type current meter from 1975 to the early 2000’s (exact date unknown), and switched to a hydroacoustic approach using a SonTek FlowTracker Handheld ADV (Acoustic Doppler Velocimeter). The changing of manual discharge equipment is likely to have little effect on the discharge measurements, with the potential for the price-type current meter to overestimate flow compared to the SonTek FlowTracker.

A stage-discharge relationship was developed using the continuous stage records and manual discharge measurements taken throughout the streamflow season to produce a rating curve. It is unclear how many rating curves in total have been developed for the Smith Creek gauging station, but it is approximated at 4 based on station analysis notes and rating curves that were obtained from public records. Although new rating curves aren't made every year, the stage-discharge values are plotted every year to confirm that the stage-discharge relationship has not deviated from the rating curve. A new rating curve is developed when streamflow deviates from the curve (for example, in the flood of 1995) and the subsequent stage-discharge values are plotted on the new curve. There are three levels of rating curves for the Smith Creek gauging station: low water (<0.8 m), medium water (0.8 - 1.7 m), and high water (>1.7 m; water levels higher than half way up the culvert). A separate rating curve was developed for conditions where streamflow overtopped the culvert (for example in 1995, rating curves showing values from 2011 and 2014 were unavailable) as the shape of the rating curve shifted under such conditions.

The control on the stage-discharge relationship was predominately identified by the WSC technicians as the culvert and road grade located immediately downstream from the gauging station. Other factors such as beaver activity and vegetation growth were also noted to affect the stage-discharge relationship (beaver activity at the culvert was only noted until 1997). Another change that had an effect on streamflow stems from the building up of the roadway that passes over the culvert in 1998 in response to the flood of 1995 where streamflow overtopped the road and threatened to wash it out. The increased road height can be seen in Figure C.2. No exact values on how much higher the road was built are available, although it was noted that the culvert was replaced with a longer one (same diameter) to account for the higher road. This allows for more water to back up behind the roadway during very high flows, which in turn can increase the maximum amount of flow owing to the increased head height allowance (Smith, 1985). In addition, the shape of the inlet was changed when it was replaced. Prior to 1998, the inlet had a sloped edge whereas afterwards it was straight (see Figure C.3). It was noted by a WSC technician in 2001 that such a change may have altered rating curve.

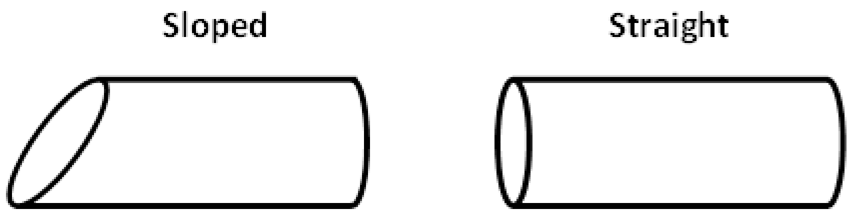

Figure C.2: Changes to the shape of the culvert at the Smith Creek gauging station. The sloped inlet existed prior to 1998 and was replaced with a straight edged inlet.

Daily discharge values are produced by the WSC using the streamflow stage and rating curves based on manual measurements and continuous stage records. The daily discharge values represent the average discharge over and entire day. As a result, the instantaneous peak daily discharge values are moderated and the annual peak discharge value is less than the instantaneous value due to the daily averaging of numbers. Table C.1 shows the instantaneous manual peak measurements compared to the corresponding daily discharge value (only four years of record with this information was available). The instantaneous discharge values are consistently higher than the daily averaged discharge values, with an average difference of ~0.645 m^3/s. The use of daily averages results in larger biases for 'flashy' peak events than more moderate, drawn out peak flows. Therefore, daily and peak discharge values slightly underrepresent the instantaneous peak daily and annual flow values. On the other hand, the influence of using the average daily streamflow values to obtain streamflow volumes would be minimal and should fairly represent volumetric values.

Table C.1: Comparing the peak discharge values between instantaneous manually measured peak flows and the average daily discharge values for the same day (Source: WSC public records).

Year	Instantaneous Peak Discharge (m^3/s)	Daily Discharge Value (m^3/s)	Difference (m^3/s)
2007	5.37	4.81	0.56
2010	3.35	3.33	0.02
2012	13.0	12.6	0.4
2013	14.6	13.0	1.6

Appendix D – Data Used

Table D.1: Annual rainfall, snowfall and total precipitation for 1942 to 2014 compiled following the methodology of this study.

Year	Rain	Snow	Total
1942	423.5	186.9	610.4
1943	246.4	164.4	410.7
1944	266.0	96.0	362.0
1945	324.2	110.0	434.2
1946	278.4	125.2	403.6
1947	309.9	114.7	424.6
1948	213.4	208.8	422.1
1949	325.4	124.9	450.3
1950	277.1	126.9	404.0
1951	301.3	170.1	471.5
1952	227.9	86.8	314.7
1953	516.7	179.0	695.7
1954	357.1	145.6	502.8
1955	310.3	152.6	462.9
1956	252.4	268.7	521.1
1957	182.7	181.8	364.5
1958	190.6	99.3	289.8
1959	350.2	167.3	517.5
1960	269.3	104.3	373.6
1961	131.0	47.4	178.4
1962	357.2	210.6	567.8
1963	429.1	53.0	482.0
1964	288.3	106.7	395.0
1965	426.1	93.4	519.4
1966	276.6	97.7	374.3
1967	252.7	204.1	456.8
1968	272.2	77.4	349.6
1969	327.4	107.4	434.9
1970	436.8	138.4	575.2
1971	431.7	102.6	534.2
1972	239.6	104.6	344.2
1973	385.7	45.0	430.7
1974	268.4	119.2	387.6
1975	388.0	97.4	485.5
1976	235.0	103.9	338.9
1977	335.5	100.4	435.9
1978	332.8	99.4	432.1
1979	216.0	111.5	327.4
1980	320.7	124.7	445.4
1981	503.4	49.8	553.2
1982	335.6	98.3	433.9
1983	341.6	145.8	487.4
1984	312.3	137.8	450.1
1985	315.1	93.3	408.4
1986	332.9	94.1	427.0
1987	323.3	88.1	411.4
1988	327.4	63.4	390.9
1989	317.5	65.5	383.1
1990	364.6	116.3	480.9
1991	377.3	154.9	532.3
1992	344.2	131.1	475.3
1993	338.9	62.9	401.8
1994	354.8	64.5	419.3
1995	388.5	185.9	574.4
1996	298.5	130.3	428.7
1997	179.4	107.0	286.4
1998	410.7	116.4	527.1
1999	426.7	94.8	521.4
2000	408.1	60.8	468.8
2001	222.7	131.3	354.0
2002	316.2	130.2	446.4
2003	201.9	149.5	351.4
2004	358.0	125.1	483.1
2005	394.9	106.8	501.7
2006	325.1	130.7	455.8
2007	321.0	85.3	406.3
2008	367.4	114.9	482.2
2009	304.8	116.8	421.6
2010	498.8	58.7	557.5
2011	378.2	126.1	504.3
2012	477.6	89.0	566.6
2013	255.0	166.0	421.0
2014	376.0	110.9	486.9

Table D.2: Annual streamflow volume, annual peak daily flow, date of annual peak daily flow, total annual duration of streamflow, runoff per unit area of the basin, and runoff ratio.

Year	Volume (dam^3)	Peak (m^3/s)	Date of Peak (Julian)	Flow Duration	Runoff (mm)	Runoff Ratio
1975	28.0	2.59	108	112	0.1	0.00
1976	7004.2	6.51	99	91	17.8	0.05
1977	224.3	0.12	91	66	0.6	0.00
1978	1120.7	0.77	90	121	2.8	0.01
1979	8913.0	6.53	112	65	22.7	0.07
1980	978.3	1.56	96	34	2.5	0.01
1981	63.8	0.06	55	50	0.2	0.00
1982	291.9	0.47	104	43	0.7	0.00
1983	2666.2	4.05	111	81	6.8	0.01
1984	366.4	0.52	86	53	0.9	0.00
1985	6606.2	4.72	105	69	16.8	0.04
1986	4978.9	5.03	90	124	12.7	0.03
1987	4529.4	5.99	95	71	11.5	0.03
1988	2322.2	2.20	94	85	5.9	0.02
1989	411.3	1.06	94	26	1.0	0.00
1990	8029.4	5.43	98	85	20.4	0.04
1991	2400.3	5.00	95	129	6.1	0.01
1992	7345.2	5.75	109	211	18.7	0.04
1993	460.2	0.50	97	108	1.2	0.00
1994	1077.8	0.67	82	209	2.7	0.01
1995	28140.3	22.60	115	192	71.5	0.12
1996	7507.3	5.94	107	63	19.1	0.04
1997	10100.8	10.60	109	78	25.7	0.09
1998	2342.6	3.26	95	130	6.0	0.01
1999	8806.2	4.82	97	167	22.4	0.04
2000	748.9	0.39	65	200	1.9	0.00
2001	16121.5	13.80	110	85	41.0	0.12
2002	547.0	1.04	106	56	1.4	0.00
2003	11242.8	11.60	102	75	28.6	0.08
2004	943.9	0.76	98	110	2.4	0.00
2005	16785.9	9.72	96	183	42.7	0.09
2006	18358.3	11.00	99	106	46.7	0.10
2007	9073.1	4.81	90	135	23.1	0.06
2008	5315.0	4.68	105	174	13.5	0.03
2009	8524.2	6.22	99	105	21.7	0.05
2010	13313.0	3.33	170	229	33.8	0.06
2011	66743.0	19.70	124	214	169.7	0.34
2012	24965.2	12.61	170	184	63.5	0.11
2013	17850.5	13.00	118	157	45.4	0.11
2014	61353.2	24.40	182	221	155.9	0.32